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SOME ASPECTS OF THE FLUVIAL GEOMORPHOLOGY
OF THE WHITEMUD BASIN, CENTRAL ALBERTA

by

ROBERT BRUCE RAINS



A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES

IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE
OF DOCTOR OF PHILOSOPHY

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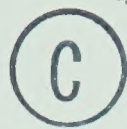
EDMONTON, ALBERTA

FALL, 1969

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UNIVERSITY OF ALBERTA
FACULTY OF GRADUATE STUDIES

The undersigned certify that they have read, and
recommend to the Faculty of Graduate Studies for accept-
ance, a thesis entitled "Some Aspects of the Fluvial
Geomorphology of the Whitemud Basin, Central Alberta,"
submitted by Robert Bruce Rains in partial fulfilment of
the requirements for the degree of Doctor of Philosophy.

ABSTRACT

Within the International Hydrological Decade programme the Whitemud basin was chosen as a catchment representative of the Park-land zone of Central Alberta. Continental glacial and glacio-lacustrine deposits, predominantly chernozemic soils, an interior continental climate, and a spring peak-flow hydrologic regime are characteristic of the area.

This study provides three sets of geomorphological information pertinent to the Whitemud basin: first, an analysis of the degree to which this catchment is morphometrically typical of the region; second, an explanatory description of landforms within the basin; and third, an interpretation of the postglacial geomorphic history of the catchment.

The morphometric analysis revealed the Whitemud basin to be typical of the area in most respects. It is markedly atypical in certain absolute and dimensional properties, for example, basin area. Interfluvial, valley and channel landforms were described mainly in terms of slopes, terrace development, channel cross-sections and riffle and pool characteristics. These data are of value for future hydrological research within the basin.

Three paired terraces in the main Whitemud valley were mapped in detail. Major stratigraphic and textural features of these deposits were described and compared. A tentative correlation of the terraces

with a sequence established for the North Saskatchewan River valley is suggested. The lower Whitemud terrace was probably formed between 8,500 and 5,000 years B.P. The middle terrace is possibly 11,000 years old, but the age of the upper unit is very uncertain. Scanty evidence of a higher aggradation surface is also present.

ACKNOWLEDGEMENTS

The valuable assistance given me by members of the examining committee throughout all stages of this investigation is gratefully acknowledged. In particular I would like to thank my supervisor, Dr. I.A. Campbell, for his unobtrusive guidance and constructive criticism of each facet of the study. Financial support provided by an International Hydrological Decade Programme grant, administered by Dr. A.H. Laycock, and the Killam Estate, is also gratefully acknowledged.

Thanks must go to Bryan Watt, Sven Djurfors, Ted Spence and Larry Stene for assistance in the field and in data-processing. The excellent technical work of Mr. J. Chesterman in reproducing the text and illustrations is more remarkable in that he was given little time in which to accomplish this. Miss J. MacLeod labored under similar time constraints in typing the dissertation. I am indeed appreciative of their patience and the quality of their work. Further technical assistance given by Mr. G. Lester, Mr. W. Jackson, Mrs. S. Kucharyshyn, Range Aerial Survey Ltd. (Calgary), and the Technical Division, Department of Lands and Forests, Government of Alberta (Edmonton), is also acknowledged.

The continuous encouragement offered by my wife, Christine, cannot go unheeded here. Her assistance in many ways, not least in typing the first draft of this dissertation under trying circumstances, is very much appreciated.

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INTRODUCTION

Fluvial geomorphology and drainage basin morphometry are fields which have received little attention in the western plains of Canada. Studies by McPherson (1966, 1968), Stalker (1968) and Westgate (1969) are exceptions, but methods of drainage basin morphometry have rarely been employed. Parry (1967) has shown, furthermore, that fluvial geomorphology in general has suffered neglect throughout the whole of Canada. Engineers have contributed to fluvial geomorphology (see: Bobey, 1965; Gray, 1964; Neill, 1964; Neill and Galay, 1967) but in most cases these studies centred on present bed and valley landforms rather than on entire basin networks.

Hydrological research has been stimulated by the International Hydrological Decade programme, within which a number of basins in participating countries were designated as specific study areas. The place of fluvial geomorphology per se in this programme is admittedly peripheral but, nevertheless, has gained active support. It was decided, therefore, that a geomorphological study of the Whitemud basin (local representative of the I.H.D. study units) would be undertaken. The primary concern was with a research design which might prove useful for subsequent hydrological work involving physiographic factors. To ensure the establishment of a wide range of geomorphic parameters the approach selected is deliberately broad and has three objectives:

CHAPTER ONE

THE STUDY AREA AND BASIC METHODOLOGY

1:1 Introduction

A rectangular area encompassing approximately 2300 square miles, bounded on the north by the North Saskatchewan River and Edmonton, south by Ponoka, east by Bittern Lake and west by Pigeon Lake, was selected (Figure 1). These boundaries were not rigorously determined but were placed to facilitate the inclusion of a reasonable number of basins, and to minimize variation in surface geology, climate, soils and vegetative cover.

1:2 Surficial Geology

Bedrock for most of the study area is the Upper Cretaceous, Edmonton Formation, consisting of brackish water bentonitic sandstones, sandy shales, bentonitic clays, and coal. However, part of the southwestern area is underlain by freshwater sandstones and soft shales of the Tertiary, Paskapoo Formation (Taylor, Mathews, and Kupsch, 1964).

At certain sites, especially where preglacial valleys existed, "Saskatchewan gravels and sands" composed almost wholly of Rocky Mountain quartzite immediately succeed the Cretaceous and Tertiary bedrock. McConnell (1885) believed these deposits to be of an age intermediate between the Miocene and Quaternary, but belonging primarily to the period immediately preceding the latter. Tyrrell (1886) thought it possible that this material derived from erosion of Miocene conglomerates. Antoniuk (1954) considered the Rocky Mountains the most likely source area.

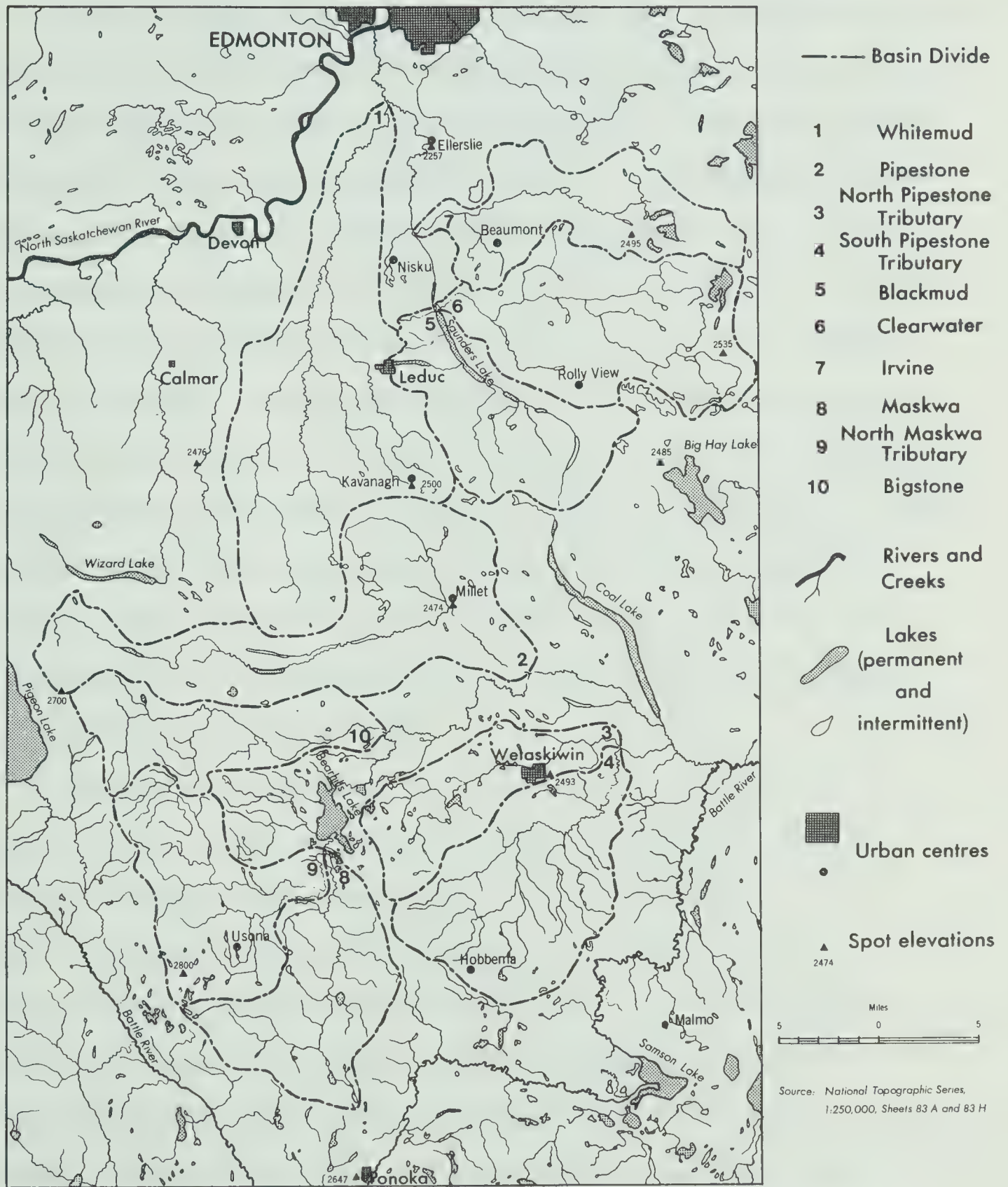


Figure 1 - The Study Area

The presence of periglacial structures in the debris led Westgate and Bayrock (1964) to conclude that it is probably of Pleistocene age but older than the classical Wisconsin.

Tyrrell (1886) found only one "boulder clay" in the Edmonton area and Rutherford (1928) showed that the general direction of the related ice sheet advance was from the north-east. Taylor (1934) and Gravenor and Bayrock (1961) supported this hypothesis. However, general agreement as to the number of distinct tills and the direction of movement of associated ice masses has not yet been attained. Duff (1951) thought that three distinct tills could be recognized but this was doubted by Bayrock and Hughes (1962), and Bayrock and Berg (1966). An intensive research programme into this and related problems has been initiated by Dr. J. Westgate, Department of Geology, University of Alberta. A number of unpublished reports which tend to substantiate the presence of at least two tills are held in that department. A short report on till differentiation in the Whitemud Creek valley (Rains, 1969) indicates that two tills are exposed there.

The surface expression of till varies from hummocky dead-ice moraine in the north-east (Bayrock and Hughes, 1962) to till plain zones of more gentle relief in the central and southern areas (Stalker, 1960). An upstanding bedrock area in the south-west is overlain by moraine forming an irregular surface. Much of the central till area is mantled by lacustrine sediments associated with the former Glacial Lake Edmonton and smaller, localized lake basins. Hughes (1958) studied the Glacial Lake Edmonton sediments in detail, part of this work being reproduced by Bayrock and Hughes (1962). Residual zones, where bedrock is at or near the surface occur in the central section, along with coarse-

textured aeolian material north and south of Wetaskiwin and south-west of Edmonton (Bowser, et. al., 1947; Bowser, et. al., 1962). Bayrock and Hughes (1962) distinguished thin deposits of early postglacial North Saskatchewan River alluvium overlying glacial Lake Edmonton sediments between Calmar and Nisku. These are not markedly different from the glacio-lacustrine material and therefore remain undifferentiated in Figure 2, which illustrates the major distributional features of surficial deposits.

1:3 Climate and Hydrology

A continental climate, designated as Dfc in the Köppen-Geiger classification (Strahler, 1969), is characteristic of the area. Detailed long-term meteorological data and analyses of these are contained in annual summary reports issued by the Dominion Public Weather Office, Edmonton.

Of greatest pertinence to this study are temperature and precipitation patterns since these are most closely associated with runoff and streamflow. Interpolation from isotherm and isohyet maps, prepared for the period 1931 to 1960 (Longley, 1967), gives the following data set.

TABLE I - CLIMATIC STATISTICS

	Degrees F.	Inches
Mean January temperature	5	
Mean April temperature	37	
Mean July temperature	61	
Mean October temperature	41	
Mean spring precipitation		6.0
Mean summer rainfall		9.0
Mean autumn precipitation		3.5
Mean annual snowfall		52.0
Total annual precipitation		18.0

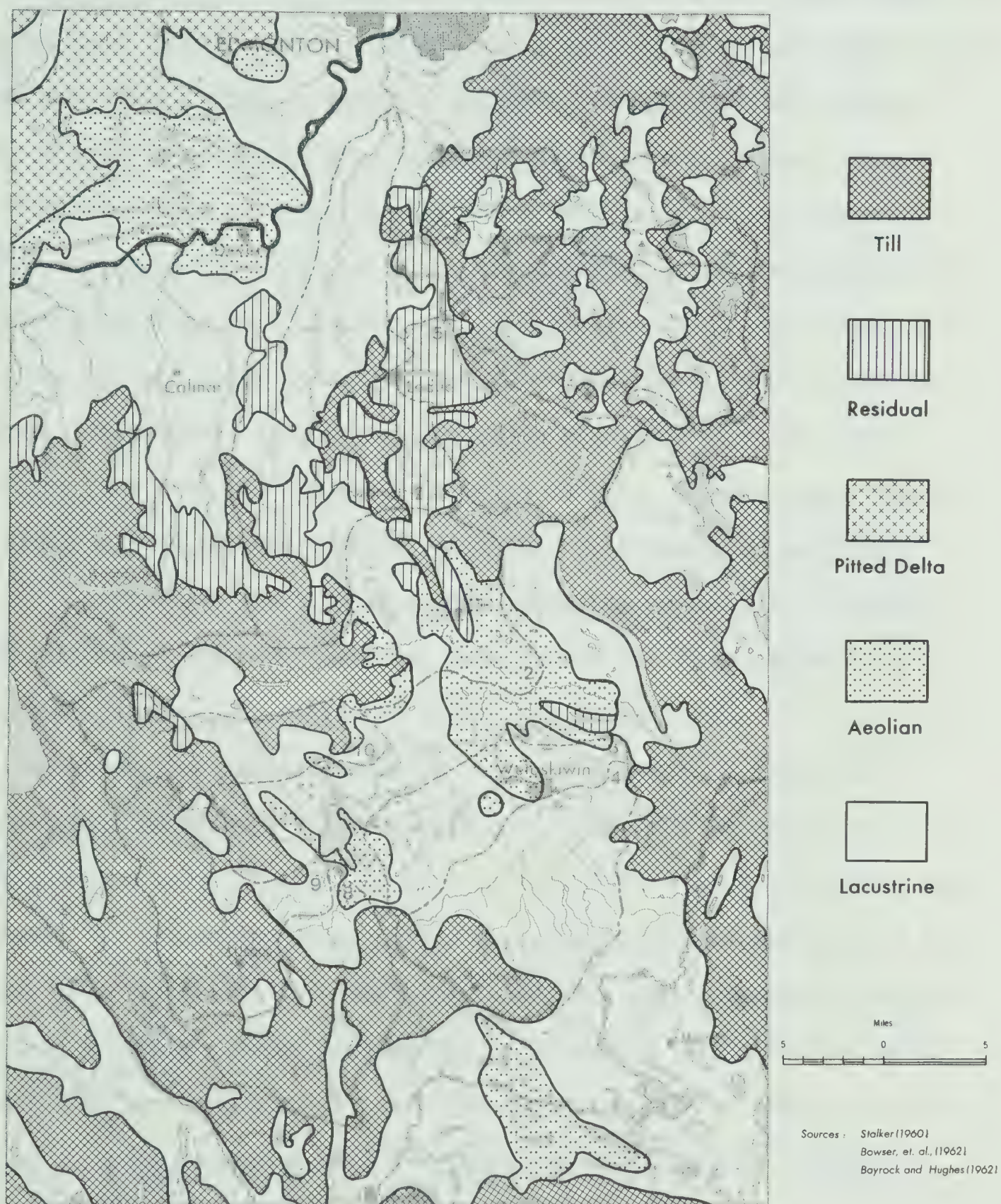


Figure 2 - Surficial Geology

Rapid rises in temperature, usually as a result of Chinook activity can accelerate erosion processes in the early spring when large and rapid runoff over a still frozen sub-soil may result in intensive erosion (Toogood and Newton, 1955). Rainfall accounts for approximately 70 per cent of the total annual precipitation¹ and is concentrated in the spring and summer seasons. Summer rains tend to be of low intensity - only one year in five is liable, on the average, to produce rainfall of over 1 inch in an hour (Bowser, et. al., 1962). Thus, the potential for erosion to take place during the period of maximum rainfall is not extreme. Variability in both precipitation and temperature for given seasons, and from year to year, is quite marked. In some years fluvial activity may be accentuated and in others, relatively insignificant.

The climate described produces a particular pattern of runoff and maximum stream flow which has been called the "snow regime of the plains" (Pardé, 1955, quoted in Sommer and Spence, 1968). Streamflow is low or may be entirely absent during winter and this condition is usually followed in late spring by a rapid rise to peak flow during maximum snow melt. Gradual decline to low levels ensues throughout the summer. In estimating water surpluses and deficiencies for part of the study area Laycock (1962, 1964) concluded that under average conditions of soil moisture storage (assumed to be 4 inches) little runoff would occur. Using meteorological data for the years 1921 to 1950 he calculated an average surplus of 1.4 inches, with a range from 0 to 5.8

¹ For local records 10 inches of snow are assumed to equal 1 inch of water.

inches, and considered that there was negligible surplus in over one-third of the years. Further, it was noted that runoff is more closely associated with snowfall and time of rain than with amount of annual precipitation. Therefore, years with relatively high total precipitation do not necessarily have a high surplus.

Very few data on specific discharges in small basins are available for the area. At most, estimates of flood-peak discharge can be obtained (Table II) but automatic stage-recorders have not yet been installed in any of the study basins.²

TABLE II - HYDROLOGIC STATISTICS

Basin ³	Area (sq.mi.)	Date	Mean velocity* (f.p.s.)	Discharge* (c.f.s.)	Maximum** stage re- corded, Apr. 1925 (c.f.s.)	Mean** monthly discharge Apr. 1925 (c.f.s.)
Whitemud	121.5	11/4/62	2.23	329	-	-
	121.5	14/4/62	2.57	812	-	-
	121.5	23/4/65	1.91	218	-	-
	121.5	26/4/67	2.59	582	-	-
Pipestone	114.0	12/4/25	-	-	1,270	286
Bigstone	260.0	13/4/25	-	-	682	183

* Source: From the files of the Water Resources Division, Govt. of Alberta, Edmonton.

** Source: Arctic and Western Hudson Bay Drainage, Water Resources Paper, 50, Dept. of the Interior, Canada, 1924-1925.

²At the present time an automatic stage-recorder is being installed in the Whitemud Creek valley.

³The basins as defined for the measurements in Table II do not terminate at the same downstream points as those defined by basin order in this study, thus the discrepancies between the areal values shown here and those presented later for the study basins.

1:4 Soils and Vegetation

Chernozemic soils, present over much of the study area, grade into grey wooded, dark grey and dark grey wooded soils on the northeastern and southwestern flanking moraines (Figure 3). A considerable part of the central area is characterized by solonetzic and solodized solonetzic soils (Bowser, et. al., 1947; Bowser et. al., 1962). In their natural state most chernozems are well-suited to resist undue erosion, in particular because of the inherently good structure and binding qualities of constituent organic matter. An exception is that of the solonetzic type in which a hard, compact, clay-rich sub-soil layer tends to develop, inhibiting infiltration capacities (Toogood and Newton, 1955). Soil moisture storage is unlikely to be uniform over such a large area as that under study. It has been noted (Laycock, 1962) that runoff may take place on clays, frozen soil, and steep slopes long before soil moisture retention capacities are reached. In general terms, though, permeability of the chernozem and podzolic soils of the area can be considered relatively good, with the exception of the solonetzic types which may be somewhat impermeable.

The natural vegetation was that of aspen parkland, transitional to boreal mixed-wood forest (Bird and Bird, 1967; La Roi, et al , 1967). Eyre (1962) described this as an extremely complicated "ecotone" - a zone of competition between two distinct plant communities. In this case the distinct plant communities were those of the true prairie grassland and the boreal mixed-wood, or poplar association (Moss, 1955). The presence of chernozem soils throughout this region led Moss (1955) to believe that for a considerable portion of Holocene time grass must

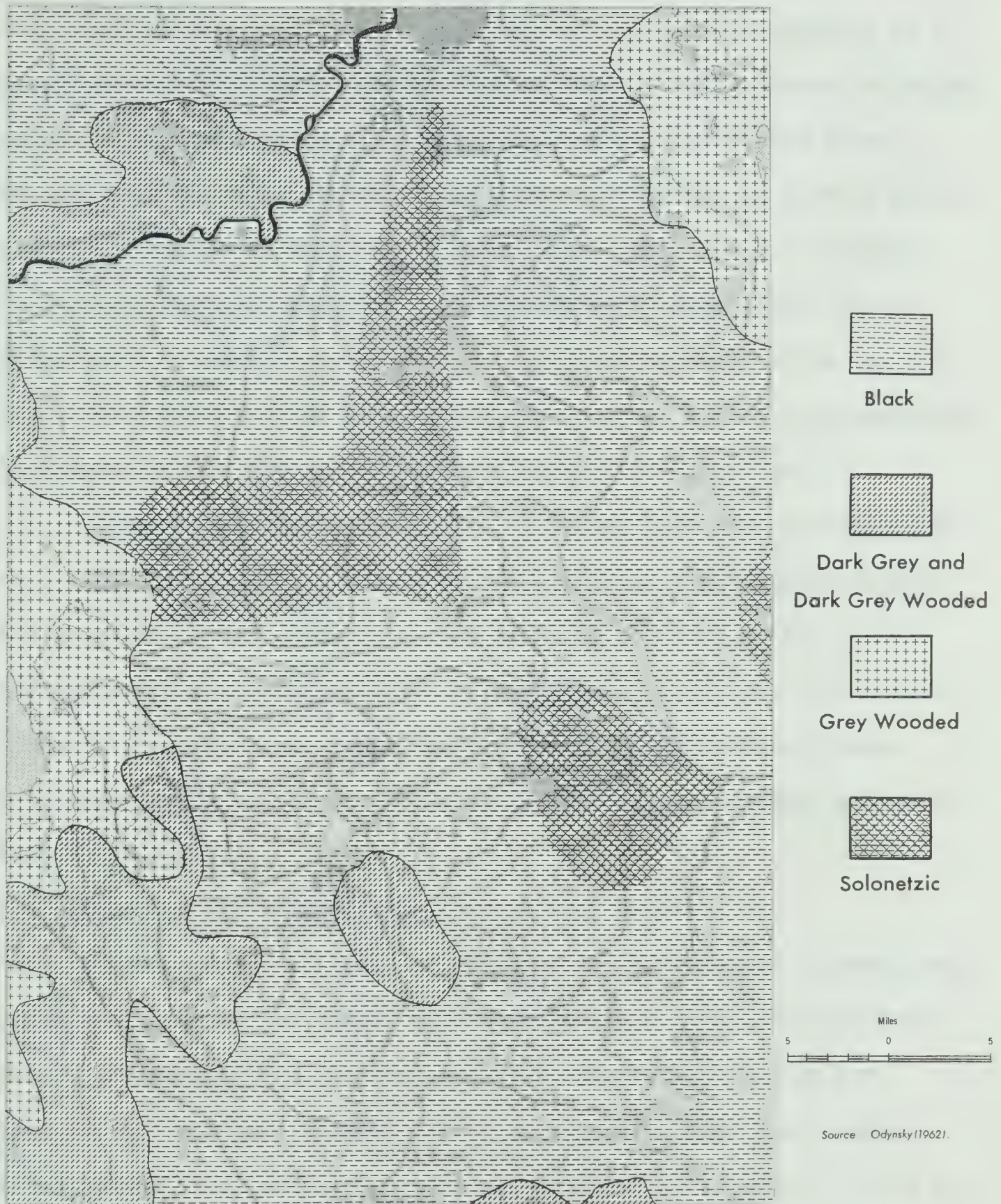


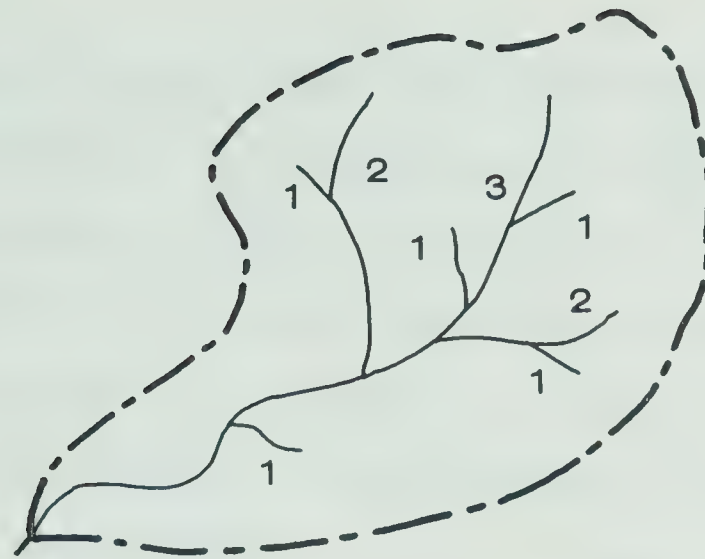
Figure 3 - Soils

have been the dominant vegetation. Further, he postulated that, as a result mainly of climatic changes and periodic burning, groves of aspen poplar have become established here in comparatively recent times. Sauer (1952) has also argued strongly for the importance of fire as an ecological factor in the prairie environment. The areas of podzolic soil on the hummocky moraines were considered by Moss (1955) to be linked with a dominant aspen poplar vegetation, along with balsam poplar. The latter thrives best on more humid sites and as a consequence the aspen is more effective in encroaching on the grassland.

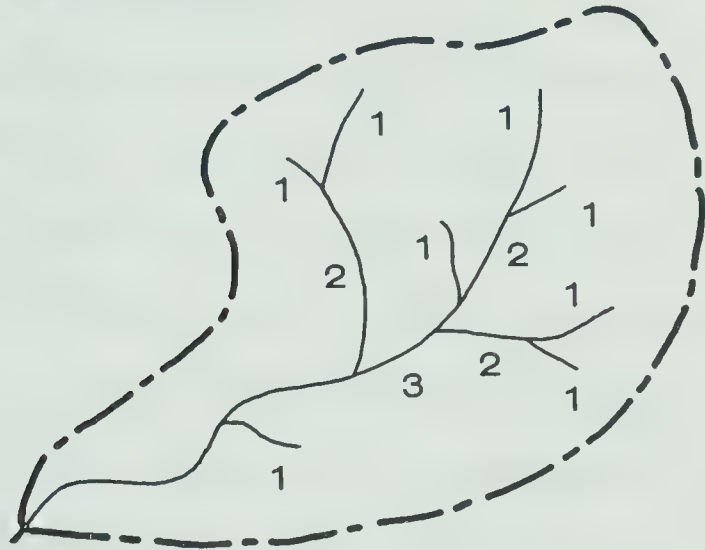
Apparently, vegetative stability cannot have been maintained in this area throughout the Holocene, and continual flux between forest and grassland plant communities has probably occurred. This has significance for any interpretation of postglacial erosional and depositional history in that changes in vegetation, from whatever cause, will have been partly responsible for stream regimen changes and debris-load fluctuations.

1:5 Method of Basin Selection

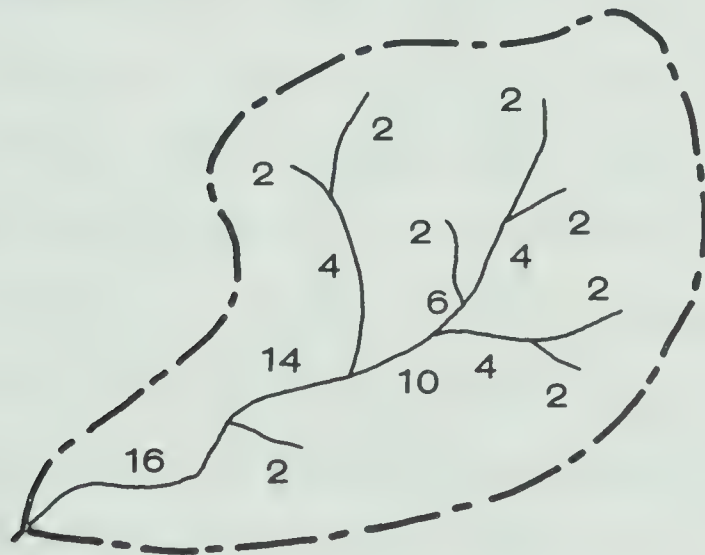
Morphometric analyses of drainage basins and their channel networks are based on the lowest common denominator of channel and basin order, a dimensionless property by which comparative studies are readily facilitated. Maxwell (1961) noted that a system of channel ordering was first proposed by Gravelius (1914) in Germany. It was not until much later in this century that the idea was developed extensively outside of Europe. Horton (1932, 1945) proposed the channel ordering system shown in Figure 4a, but Strahler (1952, 1957) preferred a modified form of the Gravelius method (Figure 4b). Melton (1959) evaluated



a after Horton



b after Strahler



c after Scheidegger

Figure 4 - Channel Ordering Systems

the mathematical basis of combinatorial analysis on which Strahler's system was developed. He concluded that this method is probably unique in excluding concepts of channel size, entrance angles and downstream direction implicitly involved in other ordering systems. Although some workers have continued to use the Horton system (see: Leopold, Wolman and Miller, 1964) Strahler's method has generally been adopted and is used in the present study.⁴

A fundamental problem in the application of stream order methods is that of scale and quality of topographic base mapping. In dealing with basins of more than a few square miles the investigator is forced to rely on the largest-scale topographic maps available for the area to be studied. In many United States investigations maps of a scale 1:24,000 and larger, produced by photogrammetric means, have been employed with apparent success when supplemented by aerial photograph checking. In some cases field checking of the drainage network was also carried out (Coates, 1956; Morisawa, 1957, 1959; Broscoe, 1959). Attempts have been made whereby additional channels were recognized from contour inflections (Strahler, 1957; Melton, 1957, Morisawa, 1959) but the method is inherently subjective (Chorley, 1958a). Obviously some combination of supplementary checks is desirable; the combination adopted will depend on time and cost considerations and on the study objectives.

⁴Recently, a revised ordering system (Scheidegger, 1965, 1968) came to the writer's attention (Figure 4c). This method removes a weakness of the Strahler method — the case of combined streams, where a segment of order u is joined by numerous lower order channels. However, since the morphometric data had already been assembled by use of the Strahler method, and, because other workers found this system adequate for comparative purposes, it was decided not to recalculate on the Scheidegger scheme.

In central Alberta the largest-scale topographic map series available for extensive areas is that at a scale of 1:50,000, with a contour interval of 25 feet (National Topographic Series, Surveys and Mapping Branch, Department of Mines and Technical Surveys). A small area surrounding the city of Edmonton is mapped at a scale of 1:25,000 but includes only an insignificant portion of the study area. The 1:50,000 series was, therefore, used as a base. Field checking of channel networks for all the basins was clearly impracticable but a comprehensive aerial photograph coverage, scale 1:31,680 (Department of Lands and Forests, Government of Alberta), enabled stereoscopic inspection of the entire expanse. It should be noted that even this method is extremely time-consuming for an area over 2000 square miles. Use of contour inflections alone as indicators of channel positions is not recommended for work in areas of continental glacial deposits. Many of the relatively sharp contour crenulations do not mark well-defined channels and, conversely, many well-defined channels do not have distinct contour inflections.

A further problem is that of channel definitions (Melton, 1957). Maxwell (1961) used various criteria to delimit channels in the field but these are not suitable when using aerial photographs as the main identification tool. Prior to the aerial photograph checking it was decided to apply the following criteria for channel recognition:

- (1) A channel must be integrated into the basin network so that concentrated runoff in that segment may eventually be carried to the main basin outlet. Therefore, well-defined channels entering depressions or sloughs but having no

surface connection with the rest of the network were excluded.

- (2) Some evidence of concentration of runoff into linear form must be observable on the scale of photograph used. This may be in the form of a well-cut channel or might be indicated, for example, by marked tonal differences on the photograph because of accentuated vegetation growth at the site.
- (3) Artificial ditches are included only where these have apparently been installed along poorly-defined natural courses. On this basis roadside ditches are excluded.
- (4) Lakes and sloughs are included in the drainage network only where a channel enters and exits from these. Where channels enter lakes at different locations selection of most probable confluence points is made by projecting azimuth lines of the relevant segments until they intersect.

Strict application of these criteria enabled relatively accurate demarcation of drainage networks throughout the area. This notwithstanding, errors of omission must be expected, especially for small first order channels. It is considered, however, that accuracy sufficient for comparative purposes has been attained and that replication of the study in similar environments of low-relief, glacial and lacustrine deposits could be achieved with similar methodology.

Application of the Strahler ordering system to the Whitemud network revealed this basin to be of fourth order. Following this, all fourth order basins falling within the study area were demarcated. Although the contour interval for the base maps is 25 feet, and local relief

is generally small, determination of the topographic divides for each basin was necessarily made from map sources. As a check on the accuracy of this delimitation a traverse was taken around the limits of the Whitemud basin during the spring of 1967. At this time the snow melt period was well under way and direction of flow in roadside ditches could be noted. In this manner a relatively accurate estimate of the divide location was obtained. Comparison with the already established boundary on the topographic base map showed only minor discrepancies amounting to less than 5 per cent of the total basin area. In total, ten fourth order basins were distinguished (Figure 1). Some of these have unnamed main channels and have thus been given titles such as North Pipestone Tributary. Each basin is usually denoted by the appropriate capital letters on diagrams. For purposes of descriptive statistical methods the ten basins may be considered as a population for the study area because all fourth order basins were included. Similarly, all selected morphometric variates in each basin were measured and analyzed. Thus, problems which may arise from sampling methodology have been eliminated. In the more detailed analysis of the Whitemud basin (Chapter Three) sampling has necessarily been employed.

CHAPTER TWO

THE MORPHOMETRY OF SELECTED BASINS

2:1 General Introduction

Description and comparison of drainage basins and their channel networks in quantitative terms arose in part from increasing dissatisfaction with the more traditional Davisian methods (see: Davis, 1909; Cotton, 1945). Over the last two decades North American geomorphologists, in extending Horton's (1932, 1945) basic work, have been especially active in developing dimensional and dimensionless quantitative measures of form in an attempt to avoid terms such as "youthful" and "early mature" which are inherently ambiguous. The increase in descriptive precision has allowed greater precision of comparative work through the use of statistical procedures. Strahler (1954) recognized early that such methods could act as a vehicle for more rigorous geomorphological explanation. Krumbein and Miller (1953) were other prominent figures in this regard. Perhaps the greatest period of production was that between the early 1950's and 1960's when Strahler and his associates at Columbia University developed many of the methods and procedures now being applied (see: Strahler, 1950, 1952a, 1952b, 1954, 1956, 1958; Coates, 1956; Broscoe, 1959; Morisawa, 1959; Maxwell, 1961). Useful summaries of most of these methods have been provided by Strahler (1964) and Zakrzewska (1967).

Morphometric measures used in this study are grouped into five sections as follows:

- (1) channel order and enumerative properties
- (2) channel length properties
- (3) basin area and area/channel length properties
- (4) basin shape properties
- and (5) basin relief properties

Figure 5 is a diagrammatic outline of the morphometric approach employed here. It may be noted that a number of sections encompass both dimensional and dimensionless measures of form. The flow-diagram is not intended to illustrate the extremely complex inter-relationships of the various morphometric measures, but rather, serves as an organizational basis for the reading of this chapter.

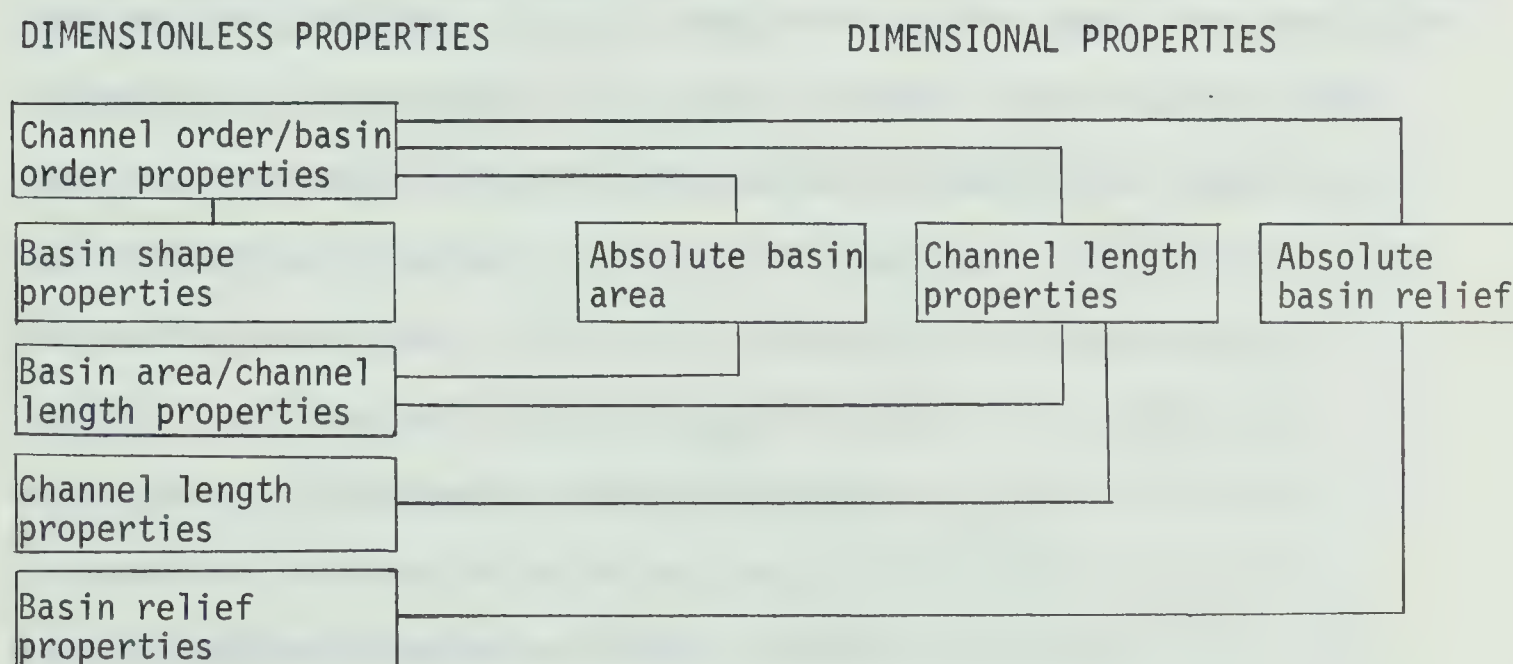


Figure 5 - An organizational framework for the description of drainage basin morphometry

2:2 Methodological Considerations

The central aim of this study is to evaluate the Whitemud basin fluvial geomorphology within the local setting to see which features of drainage development may be considered typical of the area and which may not. In dealing with a large mass of numerical data, as in morphometry, simple descriptive statistical methods may be used to, first, present the data in summarized graphical form and, second, to allow direct comparison of central tendency and dispersion measures. Inferential statistical methods go beyond simple description and comparison in estimating parameters of large populations from sample statistics. Maxwell (1961) pointed out that in studies similar to the present work, where all variates of a given property are measured (or derived) for different basins the resulting data can be considered from two viewpoints. They may be considered either as sample values from some large population of basins or as individual populations. Krumbein and Graybill (1965) made the distinction between a population of objects (the study basins) and populations of numbers (the sets of measurements of particular properties for all basins). If the selected basins are considered as a sample, two rather questionable assumptions have to be made before application of inferential statistical methods. First, it must be assumed that there exists a larger population of basins not significantly different from those studied. Second, the basins selected must be considered as a random, unbiased sample from the hypothetical population. The latter assumption is particularly dubious if all basins of a given order are selected from a study area since this is more properly a cluster sample.

Therefore, no attempt has been made here to estimate parameters of a hypothetical population. Rather, in considering the ten basins as separate sets, forming together one defined population, values measured and derived for the Whitemud basin could be compared directly with correlative values for the other basins and with some central tendency measure assumed to represent "typicality" for the entire population. An example of a similar approach is the work of Krumbein and Graybill (1965) on a number of third order basins in Arizona. To this end histograms have been used extensively to indicate the frequency distributions of the measured and derived values. Right-skewness of varying degrees is very characteristic of many frequency distributions in morphometric analysis. Logarithmic transformations are often made where inferential statistical methods require normalized distributions (see: Schumm, 1956; Broscoe, 1959). Frequency histograms for the local data range from right-skewed forms through normal to left-skewed types, depending on the property depicted. Logarithmic transformations in most cases did not normalize the distributions very well and consequently it was decided that the mode of the untransformed histograms would be taken as the best measure of central tendency. A difficulty is that comparisons made on this basis yield no numerical definition of dispersion around the central measure. In the absence of a generally applicable statistical method which would indicate the degree of divergence from the mode we are forced to rely on graphical comparison. This has been achieved merely by indicating the position of the Whitemud value on each histogram and by use of a simple ranking procedure (Section 2:9). The preceding discussion applies especially to those properties which do not

involve large populations of numbers. For example, a property such as drainage density yields only one value per basin whereas measurement of channel length for, say, all first order segments produces a large number of variates. In the latter case transformation of the data is more feasible and measures of dispersion as well as central tendency may be calculated.

2:3 Channel Order/Number Properties

2:3:1 Introduction

Strahler's method of channel ordering was adopted for this study (Section 1:5). The ten fourth order basins selected are shown in detail in Figures 6 to 15. A large contour interval was used on these maps so that details of the channel networks might not be obscured. The contours are included merely to show the general relief dimensions of the basins.

2:3:2 Property Definitions

The simple enumeration of channel segments needs no further comment but the tendency for channel numbers to approximate a semi-logarithmic distribution for any given basin requires explanation. Horton (1945, p. 291) expressed this relationship as a law of stream numbers where:

the numbers of streams of different orders in a given drainage basin tend closely to approximate an inverse geometric series in which the first term is unity and the ratio is the bifurcation ratio.

The bifurcation ratio indicates the frequency of channel branching within a network; the ratio of number of channels of one order to number of channels of the next higher order (Horton, 1945).¹ Variability in bi-

¹Note that the number of individual order bifurcation ratios must always be one less than the number of orders in the basin.

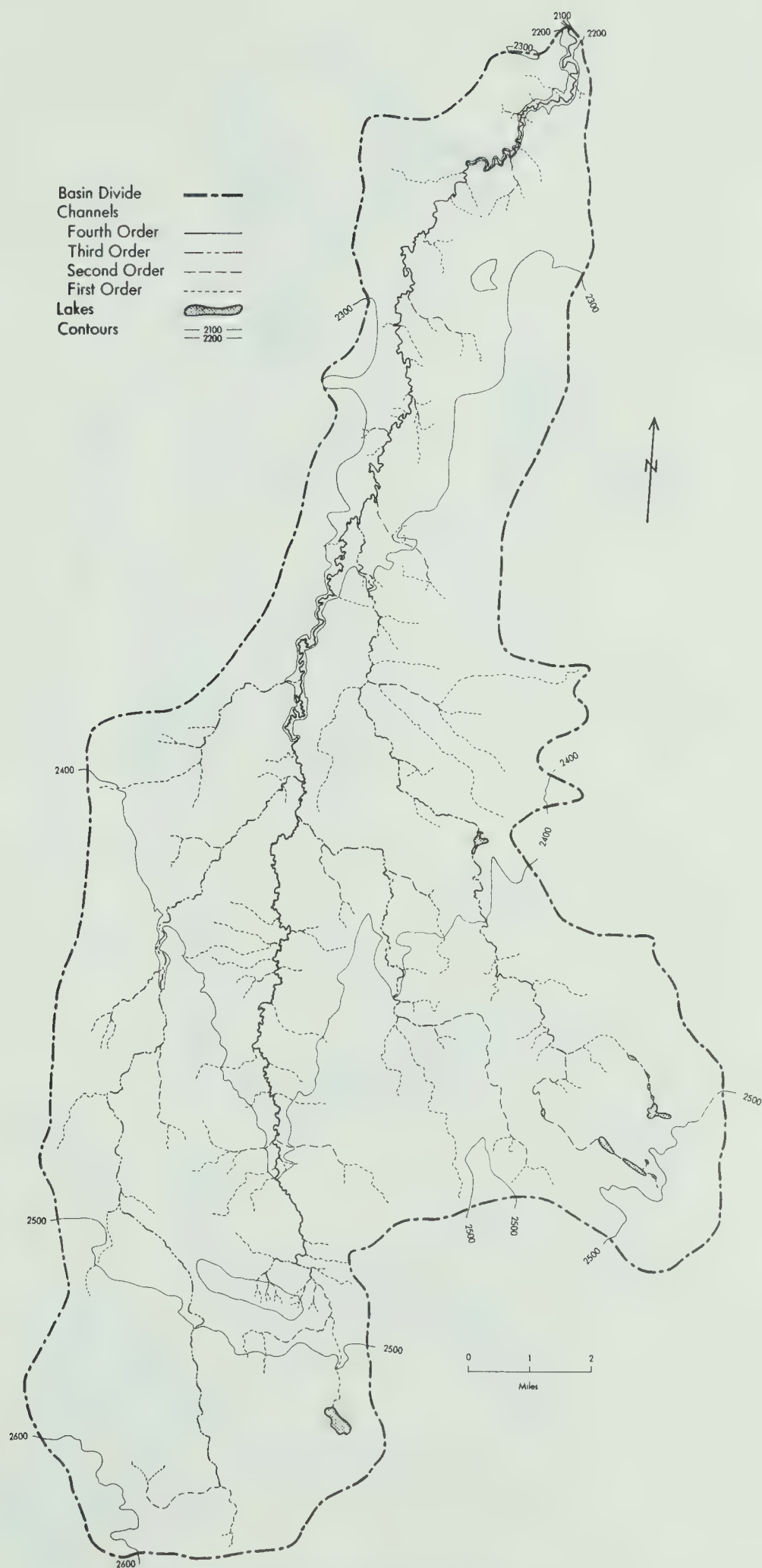


Figure 6 - The Whitemud Basin

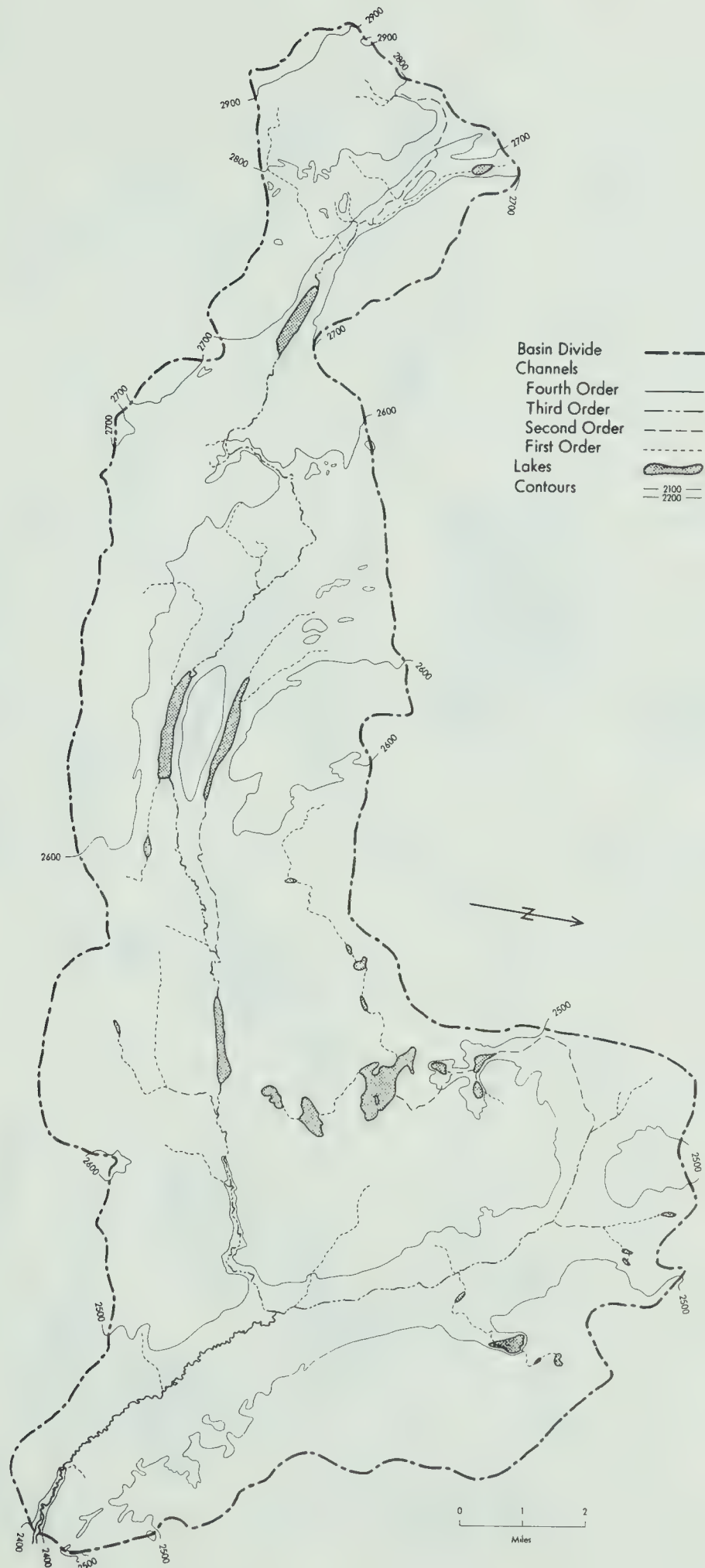


Figure 7 - The Pipestone Basin



Figure 8 - The North Pipestone Tributary Basin

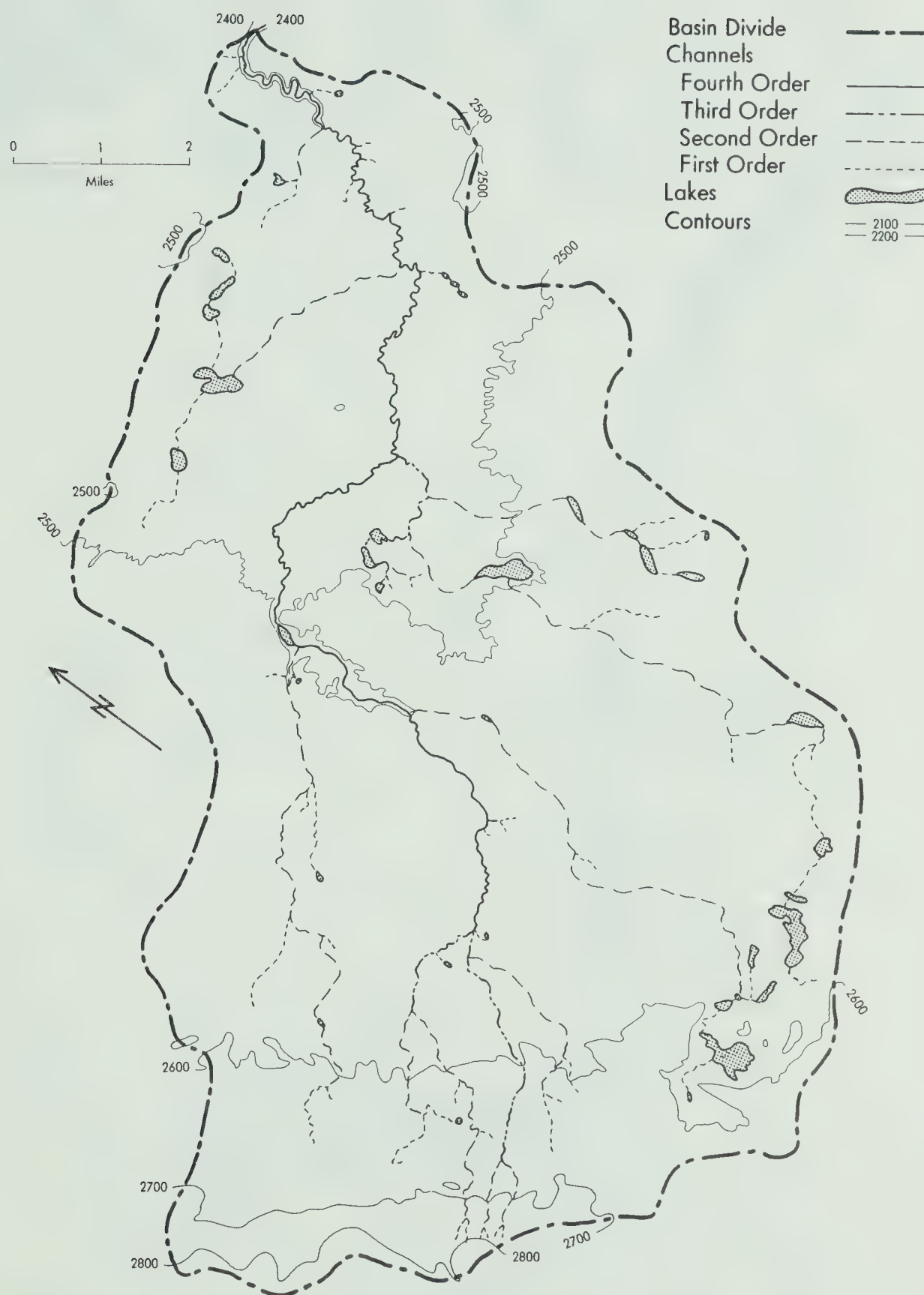


Figure 9 - The South Pipestone Tributary Basin



Figure 11 - The Clearwater Basin



Figure 12 - The Irvine Basin

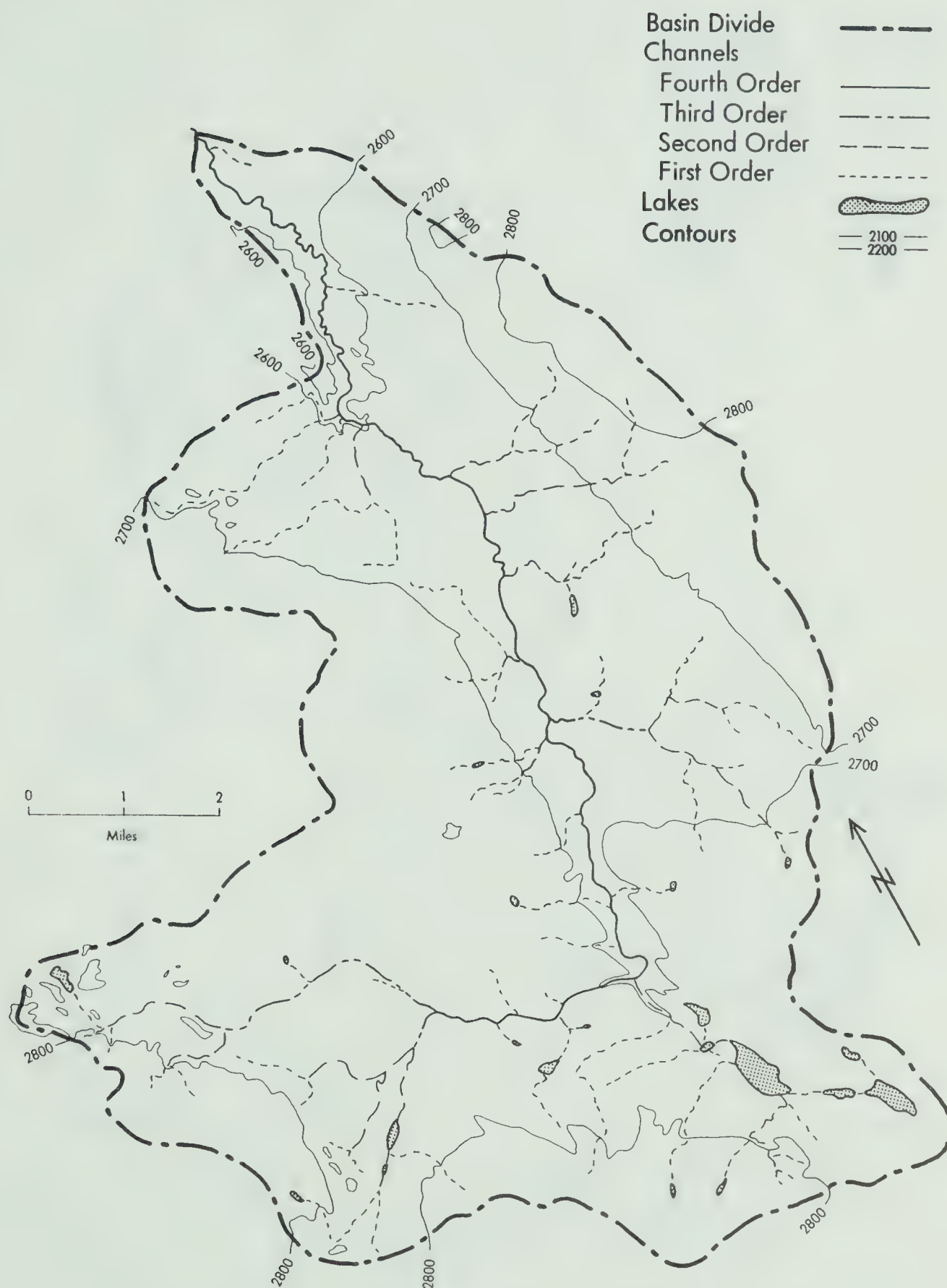


Figure 13 - The Maskwa Basin

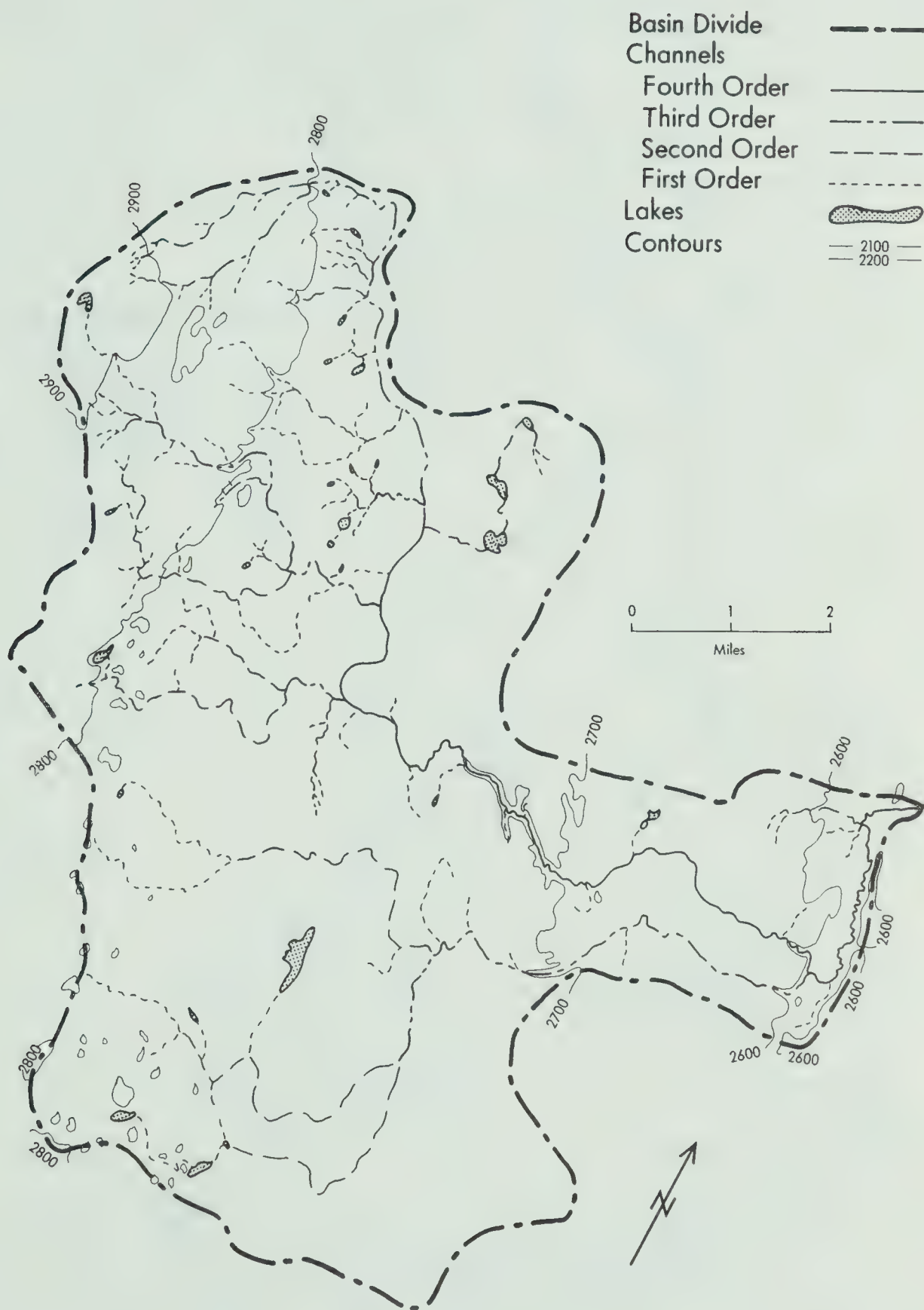


Figure 14 - The Maskwa Tributary Basin

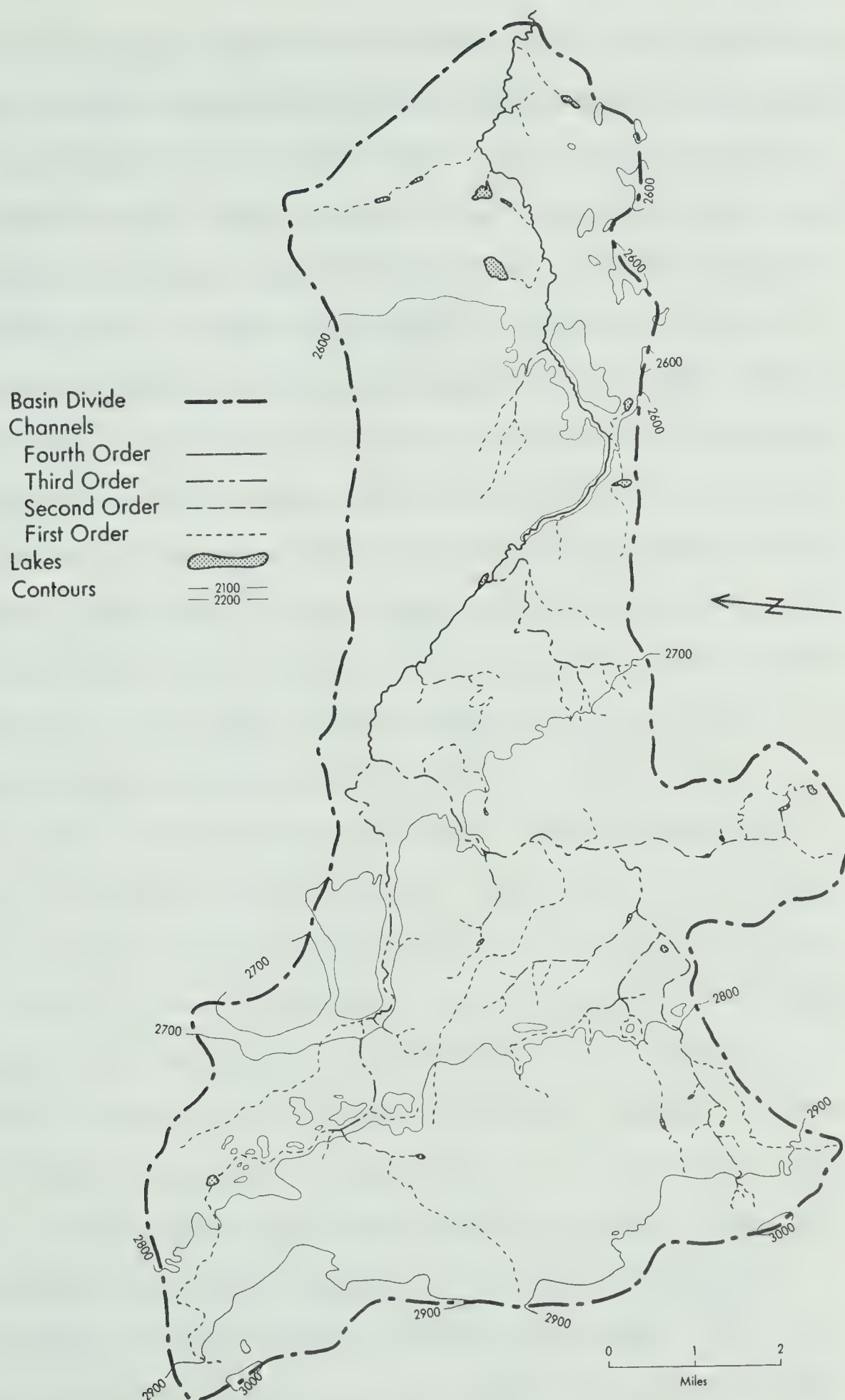


Figure 15 - The Bigstone Basin

furcation ratios usually exists for different orders within a basin but average basin ratios tend to be relatively stable, except where geologic control is very important (Strahler, 1957). Many investigators (see: Maxwell, 1961; Morisawa, 1962; Strahler, 1964) have conventionally used linear regressions to obtain best-fit lines for stream order/stream number coordinates. Following this method the bifurcation ratio is given by the coefficient describing the regression line slope. Maxwell (1961) noted that least-squares regression is not strictly justifiable in mathematical terms when using ordinal numbers, but fitting by either least-squares or by inspection yields almost identical average bifurcation ratios. Here, best-fit lines were fitted by inspection rather than by least-squares application; basin average ratios were calculated from the best-fit line slope, and the average bifurcation ratio for the total population was calculated arithmetically.

Bifurcation ratios based on the Horton ordering method differ from those derived from Strahler's system. More lower order channels are generated by the Strahler method and thus the best-fit line tends to have a steeper slope. Consequently, larger ratios will be obtained. Bowden and Wallis (1964) stated that confusion in the literature has resulted from inconsistent use of ordering systems. However, this is of greatest import when channel length properties are considered (Section 2:4). Strahler (1964) noted that bifurcation ratios based on his ordering method tend to range between 3.00 and 5.00 where geologic structures do not markedly influence the drainage pattern. He cited the case of a strike valley in an area of steeply dipping rock strata as an example where high ratios might be expected.

Some recent work of considerable interest (Shreve, 1966; Milton, 1966) indicated that Horton's law of stream numbers might be a result mainly of random processes. This would seem to show that given relatively homogeneous structure basin networks do not develop within the framework of a physical law. A forerunner to this was the generation of random-walk models by Leopold and Langbein (1962). Shreve (1966), in an analysis of 172 published sets of stream numbers, found that the most probable network order, given a number of first order channels, is that producing a geometric mean bifurcation ratio² closest to 4.00. Shreve (1966, p. 17) stated further that:

For networks with both order and number of first order streams specified, the most probable networks have the property that the bifurcation ratio of the second order streams is always close to 4 and, hence, that the bifurcation ratios respectively decrease, remain unchanged, or increase with order and the corresponding curves on the Horton diagram are respectively, concave upward, straight or concave downward according as the geometric mean bifurcation ratio is less than, equal to, or greater than 4.

Maxwell (1961) and Schumm (1956) noted that many order/number plots tend to have concave-upward profiles but they did not attempt to explain this phenomenon. Using a parabolic curve equation to define best-fit lines Shreve (1966) analyzed published data for 246 networks

²The property, geometric mean bifurcation ratio, as defined by Shreve (1966, p. 21), is essentially the same as the average bifurcation ratio when only small scatter around the best-fit line is present. It is defined as the slope of the best-fit straight line that passes through the extreme coordinates of the channel number/order graph (Horton diagram).

and found a consistent tendency for the curves to be slightly concave-upward. This means that natural drainage networks generally have a geometric mean bifurcation ratio a little less than the theoretically critical value of 4.00. Milton (1966) compared bifurcation ratios derived from random-walk model networks with those of natural basins and found that ratios for the latter tend to be lower than those for the models. He ascribed the difference to geomorphic factors, particularly drainage density and angle of channel confluence, while agreeing with Leopold and Langbein (1962) that the main element of the bifurcation ratio is that of a statistical probability function dependent on order definition. Bowden and Wallis (1964) maintained also that the semi-logarithmic relationship stems essentially from channel order definition and not from either orderly evolution or random development.

Sufficient objections have been presented in the above-mentioned studies to make Horton's concept of orderly development of drainage networks appear untenable. General agreement as to the validity or otherwise of Horton's law of stream numbers has not yet been attained, however, insofar as Woldenberg (1966) argued that this law may be justified in terms of allometric growth in a steady-state framework. Strahler (1969) endorsed this view of drainage basin development.

2:3:3 Central Alberta Data

With the above theoretical arguments in mind we may now evaluate the local data. Table III contains the number of channels in each order ($N_1 \dots N_4$), total number of channels in each basin (N_k) bifurcation ratios for each order ($R_{b2} \dots R_{b4}$), average basin bifurcation ratios ($\bar{R}_b$) and geometric mean bifurcation ratios ($\overline{GR}_b$).

TABLE III - CHANNEL NUMBERS AND BIFURCATION RATIOS

Basin	N_4	N_3	N_2	N_1	N_k	R_{b2}	R_{b3}	R_{b4}	$\bar{R}_b$	$\overline{GR}_b$
Whitemud	1	6	38	145	190	3.82	6.33	6.00	5.37	4.90
Pipestone	1	2	8	32	43	4.00	4.00	2.00	3.31	3.16
N.P.T.	1	2	7	31	41	4.43	3.50	2.00	3.24	3.09
S.P.T.	1	3	14	54	72	3.86	4.67	3.00	3.98	3.72
Blackmud	1	3	18	75	97	4.17	6.00	3.00	4.27	4.27
Clearwater	1	4	19	85	109	4.47	4.75	4.00	4.37	4.37
Irvine	1	3	11	41	56	3.73	3.67	3.00	3.55	3.55
Maskwa	1	3	16	62	82	3.89	5.33	3.00	3.98	3.98
M.T.	1	5	24	100	130	4.17	4.80	5.00	4.90	4.90
Bigstone	1	3	14	62	80	4.43	4.67	3.00	3.98	3.98
Averages	1	3	17	69	90	4.08	4.97	3.40	4.10	3.99

Plots of logarithm (to the base ten) of channel numbers against order (Figure 16) show at most only moderate scatter. A slight up-concavity of the coordinates is also evident for many of the local study basins, except for those with $\overline{GR}_b$ values much larger than 4 (Table III and Figure 16). It may be noted that these plots, central to Horton's law of stream numbers, are relatively insensitive to quite large fluctuations in channel numbers of the lower orders. For example, a basin such as Whitemud could feasibly have twenty more first order channels added without altering the graph significantly. This serves to emphasize that the relationship is not particularly valuable as a descriptive and comparative measure.

An inspection of Figure 17 a-c will show that, in absolute terms, the Whitemud basin has three times as many first order channels, two and one half as many second order and twice the number of third order channels than the modal (typical) values. Since the Whitemud basin is the largest

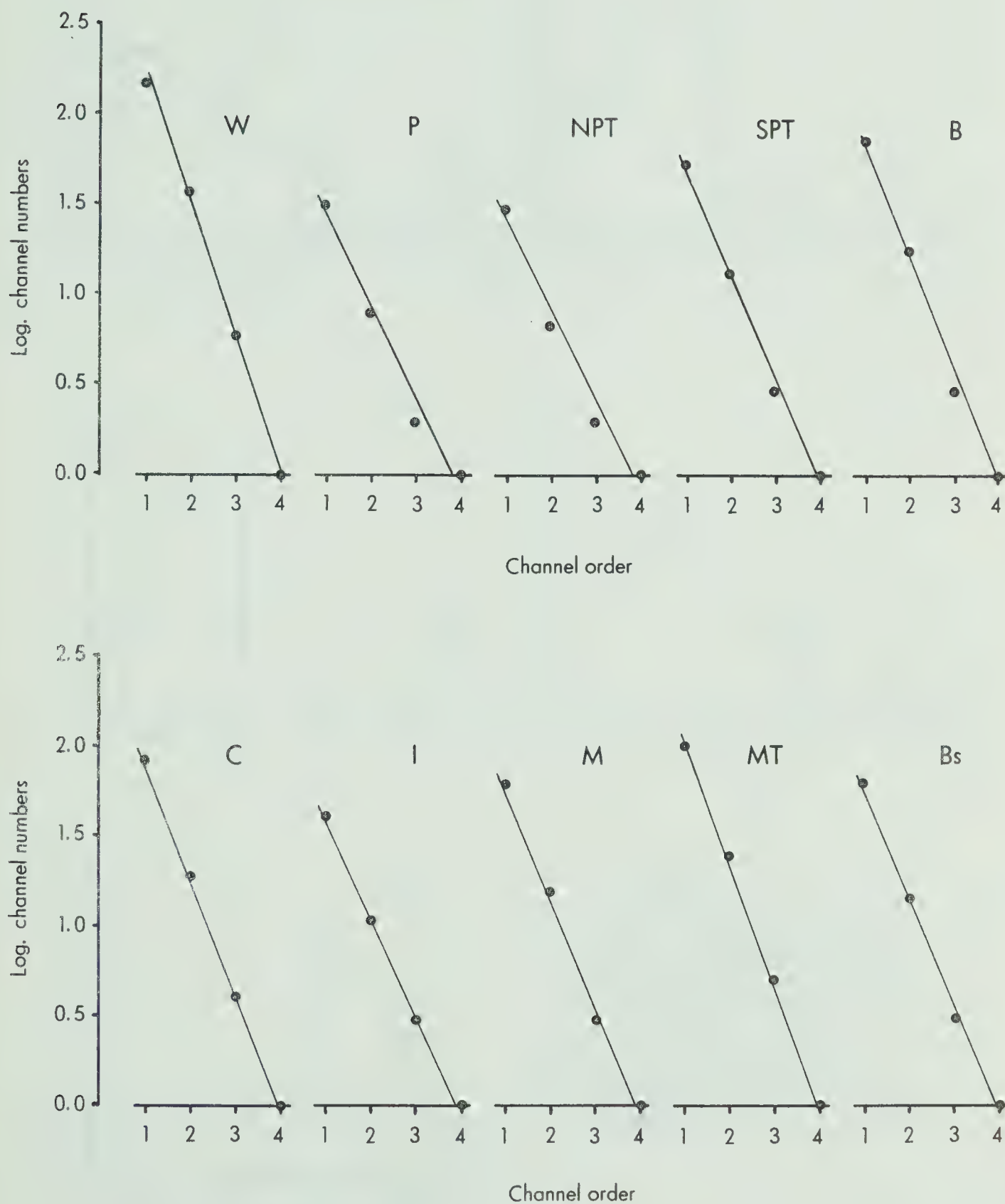


Figure 16 - Logarithm of Channel Number for Each Order

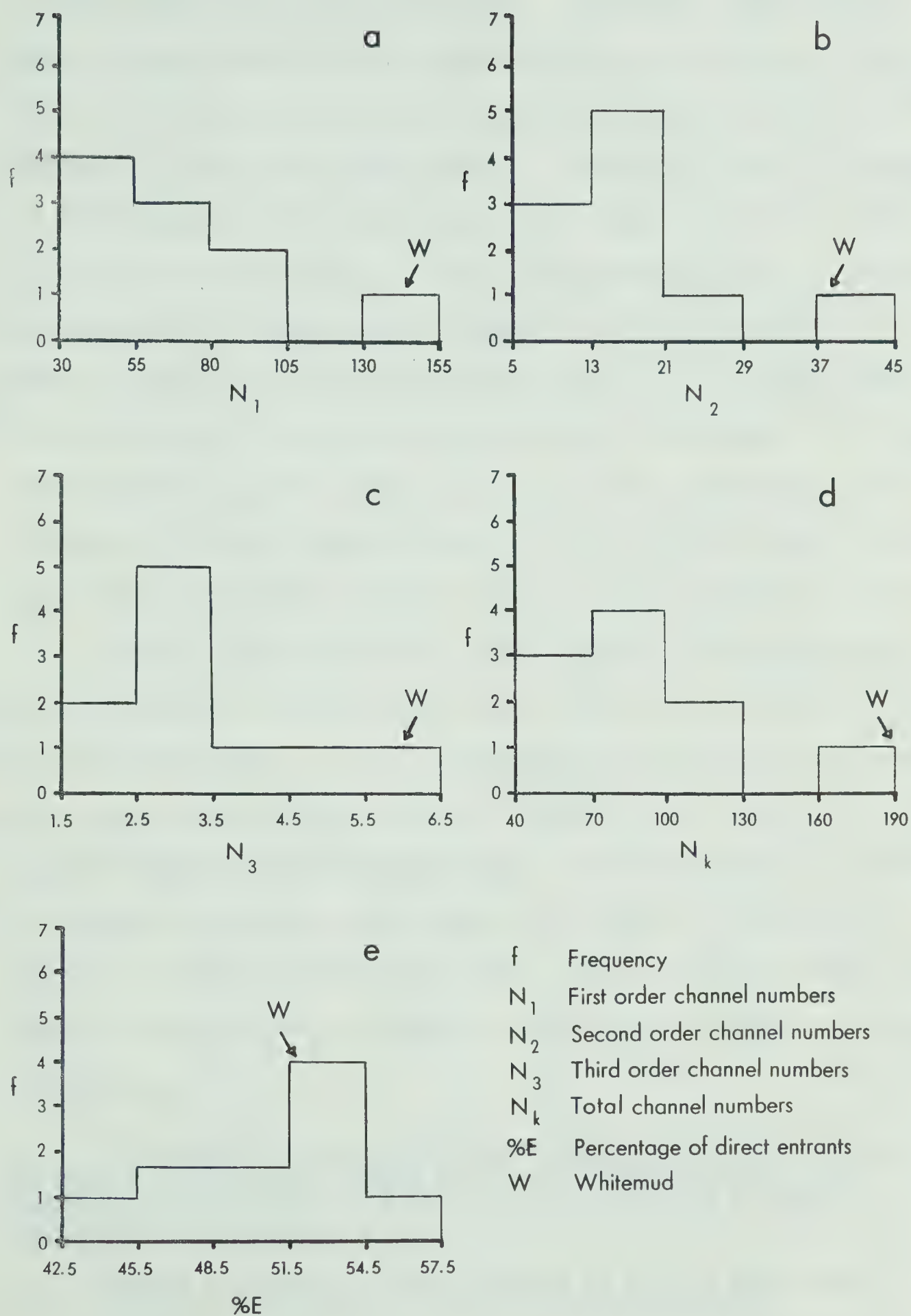


Figure 17 - Channel Number/order Values and Percentage of Direct Entrants

of those studied this is not unexpected. Therefore, these absolute values are more meaningful when combined with area as ratios (Section 2:5). It follows, also, that the Whitemud bifurcation ratios are rather atypical of local fourth order basins.³ Bifurcation ratios of the second order segments for all ten basins range from 3.73 to 4.47 (Table III), with an average value of 4.08. This supports Shreve's (1966) observation that the most probable geometric ratio for this order is 4.00. Table III shows that average bifurcation ratios for the basins range from 3.24 to 5.37, the latter value being that of Whitemud. If in fact strong geologic control produces an average ratio greater than 5.00 (Strahler, 1957) such control(s) may well exist for this basin. However, Shreve (1966) would consider this as simply a matter of chance.

Leopold, Wolman and Miller (1964) observed that approximately half the streams of a given order enter directly into channels of two or more orders higher. Since they employed the Horton ordering system percentages calculated on the basis of Strahler order must differ. For a given channel network the percentage of direct entrants⁴ will tend to be less for the Strahler order than for the Horton. The reason is obvious — with the latter method higher order channels are given greater length by the extension of orders to channel heads (Figure 4a). As a

³Histograms of bifurcation ratios were not included because the most definitive ones (those of third and second order, but especially the latter) have almost even distributions for the population of ten basins.

⁴Direct entrants are those segments of a given order which do not combine to initiate a segment of next higher order.

result, for the given network, small low order tributaries have a greater chance of entering higher order channels. Table IV contains the relevant data for the ten local basins under the following categories:

- (1) percentage of first order channels entering directly into third and fourth order channels ($u_1: u_{\geq 3}$)
- (2) percentage of first order channels entering directly into second, third, and fourth order channels ($u_1: u_{\geq 2}$)
- (3) percentage of second order channels entering directly into fourth order channels ($u_2: u_4$)
- (4) percentage of second order channels entering directly into third and fourth order channels ($u_2: u_{\geq 3}$)
- and (5) percentage of all direct entrants ($u_k: N_k$).⁵

The raw data from which the percentages were calculated are given in parentheses.

TABLE IV - DIRECT ENTRANTS (%)

Basin	$u_1: u_{\geq 3}$	$u_1: u_{\geq 2}$	$u_2: u_4$	$u_2: u_{\geq 3}$	$u_k: N_k$
Whitemud	35(51:145)	48(69:145)	37(14:38)	68(26:38)	52(99:190)
Pipestone	38(12:32)	50(16:32)	0 (0:8)	50 (4:8)	47(20:43)
N.P.T.	26(8:31)	52(16:31)	14 (1:7)	43 (3:7)	46(19:41)
S.P.T.	22(12:54)	48(26:54)	36(5:14)	57(8:14)	49(35:72)
Blackmud	32(24:75)	52(39:75)	56(10:18)	67(12:18)	54(52:97)
Clearwater	36(31:85)	55(47:85)	21(4:19)	58(11:19)	55(60:109)
Irvine	29(12:41)	49(19:41)	27(3:11)	45(5:11)	45(25:56)
Maskwa	21(13:62)	48(30:62)	56(9:16)	63(10:16)	50(41:82)
M.T.	25(25:100)	52(52:100)	13(3:24)	51(14:24)	53(69:130)
Bigstone	27(17:62)	55(34:62)	21(3:14)	50(7:14)	53(42:80)
Average	30(205:687)	51(348:687)	31(52:169)	59(100:169)	51(462:900)

⁵Also added to u_k are third order segments which enter directly into fourth order channels.

The decrease of direct entrants into channels two or more orders higher ($u_1: u_{\leq 3}$ and $u_2: u_4$) from about 50 per cent noted by Leopold, Wolman and Miller (1964) to near 30 per cent (Table IV) probably results from the different ordering systems used. Perhaps of greater interest is the fact that, on the average, approximately 50 per cent of all channel segments directly enter channels of at least one order higher for the central Alberta examples. Figure 17e indicates that the distribution of $u_k: N_k$ values for the ten basins is left-skewed but that the Whitemud value falls in the modal class.

2:3:4 Summary

It has been shown above that the Whitemud basin is, in terms of absolute lower order channel numbers and bifurcation ratios, atypical of local fourth order drainage basins. Whether this indicates that orderly development of the Whitemud basin has been encouraged by peculiar geological factors or, alternatively, that the high values of these properties occur by chance is debatable. Further consideration of these arguments will be postponed until the remaining properties have been presented and discussed.

Variability in percentages of direct entrants is greatest for individual orders, explained largely by the changing bases of percentage calculation. Thus, the greatest aberrations occur where numbers of channel segments are small. Average values for each basin appear particularly stable, however, and the Whitemud basin certainly does not show any significant abnormality.

2:4 Channel Length Properties

2:4:1 Introduction

Accurate measurement of channel lengths for any drainage network is highly dependent on the application of consistent criteria for the recognition of channels (Section 1:5). Networks shown on most topographic maps tend to have inconsistencies in the inclusion and exclusion of channels at a given map scale. The problem is compounded by inevitable cartographic variability when a large number of personnel are involved in the map-series compilation. However, where careful checking from aerial photographs is carried out by one investigator, as in this study, cartographic bias may be considerably reduced.

Since the first order channels for any network are usually the smallest and least well-defined it is to be expected that the greatest variability in both recognition and length measurement will accompany these. Comments have often been made in the literature regarding difficulties in defining the headward limit of such low order channels (see: Melton, 1957; Morisawa, 1957; Maxwell, 1961). It is virtually useless to apply field criteria to this problem in the local area. Widespread cultivation obliterates a large amount of evidence of recent incision, especially that of first order channels. In addition, many of the channels as recognized from stereoscopic inspection of aerial photographs have the form of shallow depressions with little clear incision. As a consequence the first order channel heads were delimited solely from the aerial photographs. It is recognized that consistency in demarcation will have been affected by such factors as photographic tonal differences and varying degrees of concentration on the part of the investigator.

However, it is reasonable to assume that such errors are in part self-compensating. In this regard it is reassuring that first order channels yield the largest number of channel length variates for each basin.

It is common practice to use a chartometer for measuring lengths but the writer encountered considerable problems with this for the more sinuous channels. Although great care was taken to measure accurately, errors of up to 30 per cent of actual channel length resulted in some cases. Of necessity, then, channel length measurements were laboriously repeated by use of a graduated scale, read to 0.05 of a mile, ensuring a high degree of accuracy.

2:4:2 Property Definitions

Channel segment length data may be conveniently summarized by calculating total length for the whole basin, the longest channel segment lengths in each order and the average length in each order. Horton (1945) proposed that channels should be fully extended to the related basin divide along a path orthogonal to the contours. This he termed "mesh-length." On inspection of Figures 6 to 15 it will be realized that such extension is not warranted for local drainage networks. In many instances the heads of first order channels are separated from divides by distances exceeding the length of those channels. Basin divides are often located in almost flat terrain and extension of channels across such areas would not represent reality. Indeed, the peripheral zone of each basin not yet penetrated by the headward retreat of first order channels is itself of prime interest. It is strongly recommended, therefore, that morphometric studies in low-relief regions should not include "mesh-length" as a measurement criterion unless well-defined

channel heads are generally in close proximity to basin boundaries. A further requirement in this regard is the availability of maps having a small contour interval.

Earlier mention of the law of stream numbers leads now to the corollary law of stream lengths, which has been defined as follows (Horton, 1945, p. 291):

The average lengths of streams of each of the different orders in a drainage basin tend closely to approximate a direct geometric series in which the first term is the average length of streams of the 1st order.

Again, confusion has occurred in the past when testing of this relationship was based on Strahler channel ordering and not on Horton's method. Broscoe (1959) suggested that plotting of cumulative mean channel lengths against order should accompany the Strahler ordering system. At first unaware of Broscoe's suggestion Bowden and Wallis (1964) reached the same conclusion from a study of sixty basins in California. They then tested the law of stream lengths, using the Broscoe (1959) definition of cumulative mean length, for sixty-six watersheds from eight physiographic provinces and achieved good semi-logarithmic fits throughout on this basis. They also indicated that this relationship is essentially a statistical one, analogous to the law of stream numbers (Shreve, 1966; Milton, 1966). However, they did not reject this as an analytical measure because they considered that physical properties of drainage networks are indicated by the regression-line slopes. Calculation of the regression-line slope yields the length ratio, or the cumulative mean length of one order to the cumulative

mean length of the next lower order.⁶

A considerable amount of research has been directed toward analysis of channel and valley sinuosity (see: Bates, 1939; Leopold and Maddock, 1953; Dury, 1954; Leopold and Wolman, 1957, 1960; Schumm, 1960, 1967a; Mueller, 1968). A comprehensive review and evaluation of various sinuosity indices was presented by the last-named author. In this article Mueller (1968, p. 372) stated:

....if topography is not accounted for in existing indexes, all indexes are somewhat meaningless when the measured streams flow across a topography not created by the streams themselves. This point is critical, especially when examining the many rivers found in freshly glaciated (Wisconsin Stage) areas.

He distinguished two types of sinuosity, hydraulic and topographic; the former being related to open channel hydraulics, the latter resulting primarily from effects of initial topographic irregularities. He considered the efficacy of hydraulic factors in the production of sinuosity to be at a minimum during the pre-floodplain formation stage of channel development. Thereafter, with continuing floodplain extension hydraulic sinuosity should increase in importance. However, he maintained that topographic sinuosity will persist throughout drainage development as long as the valley remains irregular. Although mainly concerned with the analysis of river reaches, Mueller's (1968) methodology would seem

⁶As usually defined, length ratios are ratios of the mean length of channels in a given order to the mean length of channels of the next lower order. Following Strahler (1964), in this study, a length ratio for each basin was taken to be the antilogarithm of the regression coefficient, b , for the equation $y = a + bx$ describing the regression lines in Figure 18.

to be of great value in comparing entire drainage networks. This may be expedited simply by joining with straight lines all channel heads and channel mouths of segments in each order. Values of the resulting straight-line distances may be compared with the natural lengths in the form of ratios. Data thus obtained could be treated by conventional methods of frequency distribution analysis with the appropriate transformation. Here, simple comparisons of average sinuosity ratios for each order will suffice.

A measure termed "length of overland flow" which may be used to describe either the average or maximum length of non-channel surface runoff in first order basins (Horton, 1945; Broscoe, 1959; Strahler, 1964) was not included in this study. Extreme difficulties were encountered in attempting to delimit first order basins within the fourth order units studied. In the absence of topographic maps having a very small contour interval, each basin would have to be demarcated individually in the field — an impossible task for such a large study area.

2:4:3 Central Alberta Data

Values of total channel length for each order ($L_1 \dots L_4$), total channel length for each basin (L_k), mean channel length for each order ($\bar{L}_1 \dots \bar{L}_3$) and basin ($\bar{L}_k$), maximum channel segment length for each order ($L_{\max 1} \dots L_{\max 3}$)⁷ and channel length ratios (R_L) are contained in Table V. For ease of graphical representation logarithms (to the base ten) of the cumulative mean lengths were utilized. This necessitated the conversion of arithmetic mean channel lengths (miles) into one hundredths of miles, to allow use of only positive logarithms.

⁷The values of L_4 , $\bar{L}_4$ and $L_{\max 4}$ are obviously identical and are therefore indicated only as L_4 .

TABLE V - CHANNEL LENGTH VALUES

Basin	----- (in units of miles) -----					R_L
	L_1	L_2	L_3	L_4	L_k	
Whitemud	88.25	30.45	36.75	47.00	202.45	4.68
Pipestone	45.75	16.50	28.00	6.20	96.45	2.95
N.P.T.	19.60	11.10	7.40	8.00	46.10	2.75
S.P.T.	26.35	33.45	6.40	16.25	82.45	3.39
Blackmud	49.40	20.30	6.15	6.80	82.65	2.40
Clearwater	46.35	31.85	20.50	10.60	109.30	3.16
Irvine	26.40	14.30	6.55	11.30	58.55	2.63
Maskwa	46.55	17.65	4.70	14.40	83.30	2.63
M.T.	40.35	28.50	17.55	12.00	98.40	3.39
Bigstone	50.75	14.35	13.85	10.95	89.90	2.82

Basin	----- (in units of miles) -----					
	$\bar{L}_1$	$\bar{L}_2$	$\bar{L}_3$	$L_{\max 1}$	$L_{\max 2}$	$L_{\max 3}$
Whitemud	0.61	0.80	6.13	3.70	3.80	15.00
Pipestone	1.43	2.06	14.00	6.50	5.30	21.60
N.P.T.	0.63	1.59	3.70	2.20	5.25	6.20
S.P.T.	0.49	2.39	2.13	3.40	6.45	2.90
Blackmud	0.66	1.13	2.05	2.60	3.70	2.80
Clearwater	0.55	1.68	5.13	1.95	3.80	8.45
Irvine	0.64	1.30	2.18	2.75	4.90	4.20
Maskwa	0.75	1.10	1.57	2.50	3.25	2.65
M.T.	0.40	1.19	3.51	1.80	3.40	5.95
Bigstone	0.82	1.03	4.62	4.70	3.90	6.10

Most of the study basins appear to conform to the modified law of stream lengths, with moderate scatter around the regression line of logarithm of cumulative mean channel length on order (Figure 18). The upconcavity of the Whitemud coordinates results primarily from the exceptionally long fourth order channel in this basin (Table V).

As has been noted by Milton (1966) the greatest deviations from best-fit straight lines might be expected for the order with the smallest number of channels, in this case, one channel. Figure 18 and Table V

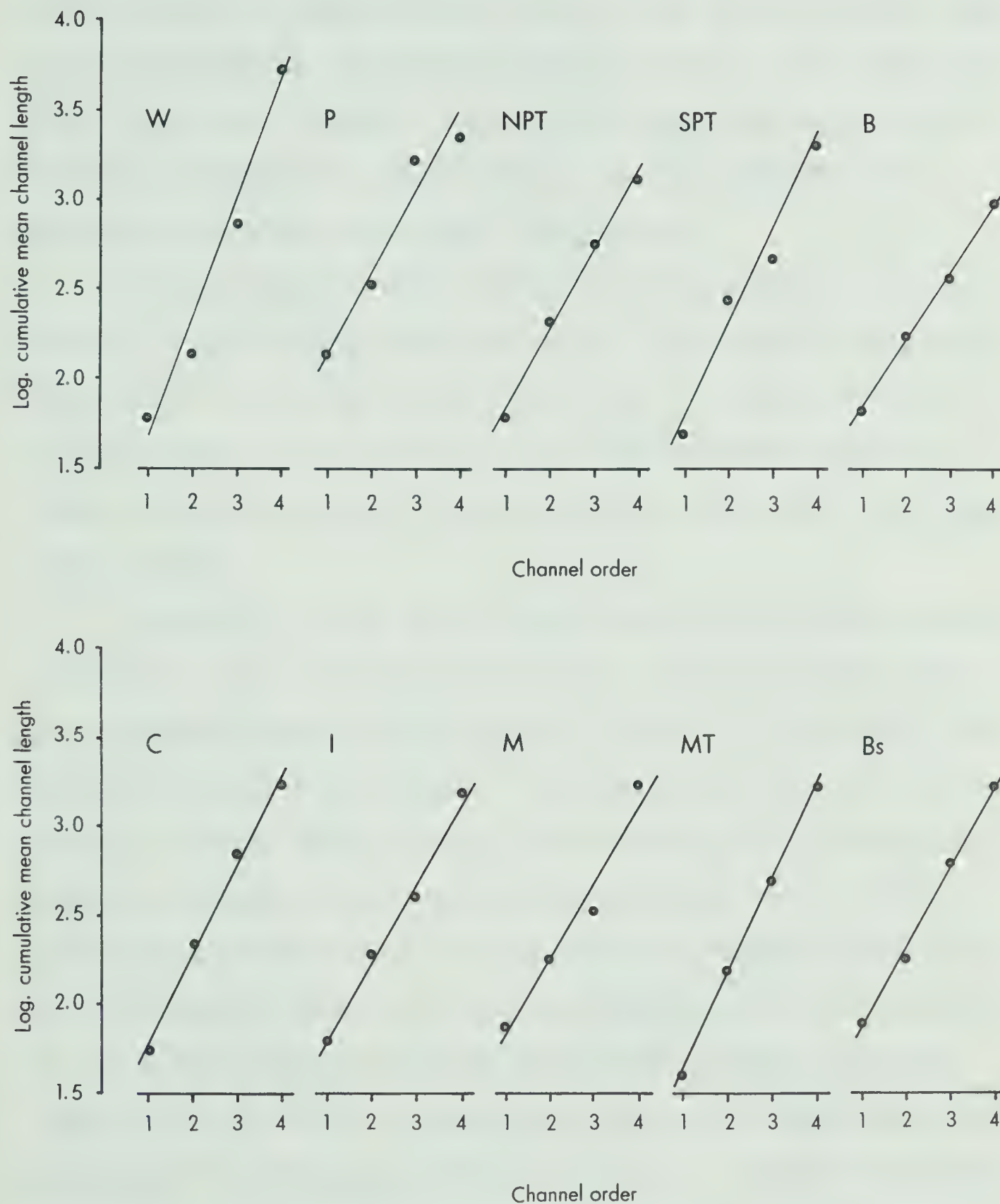


Figure 18 - Logarithm of Cumulative Mean Channel Length for Each Order

show that, in relation to other values in each individual set, the Pipestone basin has an abnormally short fourth order channel and very long third order channels; the Maskwa catchment contains a rather long main channel; the South Pipestone Tributary basin shows some degree of abnormality throughout all orders and the remaining networks have relatively consistent mean channel length values.

Average lengths of first, and second order channels, and of all channels, in the Whitemud basin tend to fall in or near the respective modal classes for the ten basins (Figure 19). The length of the main Whitemud channel (Figure 19d) may be considered markedly atypical of local fourth order basins. Third order channels also tend to be longer than usual here.

The question arises whether the arithmetic mean length values are reliable for inter-basin comparisons. Many investigators have noted that arithmetic distributions of channel lengths for given orders tend to have pronounced right-skewness. Transformation by plotting the logarithms of channel length against their frequency of occurrence usually produces a histogram closely approximating a normal curve. Following this, descriptive measures such as the mean and standard deviation may be calculated and compared, whereas such measures may be of limited use for the skewed arithmetic distribution (Chorley, 1966). All channel length values for first and second orders were accordingly transformed⁸ and grouped into histograms (Figures 20 and 21). Frequencies were converted to percentage frequencies so that the area of each histogram would remain equal, regardless of the number of variates classified

⁸Again, the measured values were first converted to one hundredths of miles to allow use of only positive logarithms.

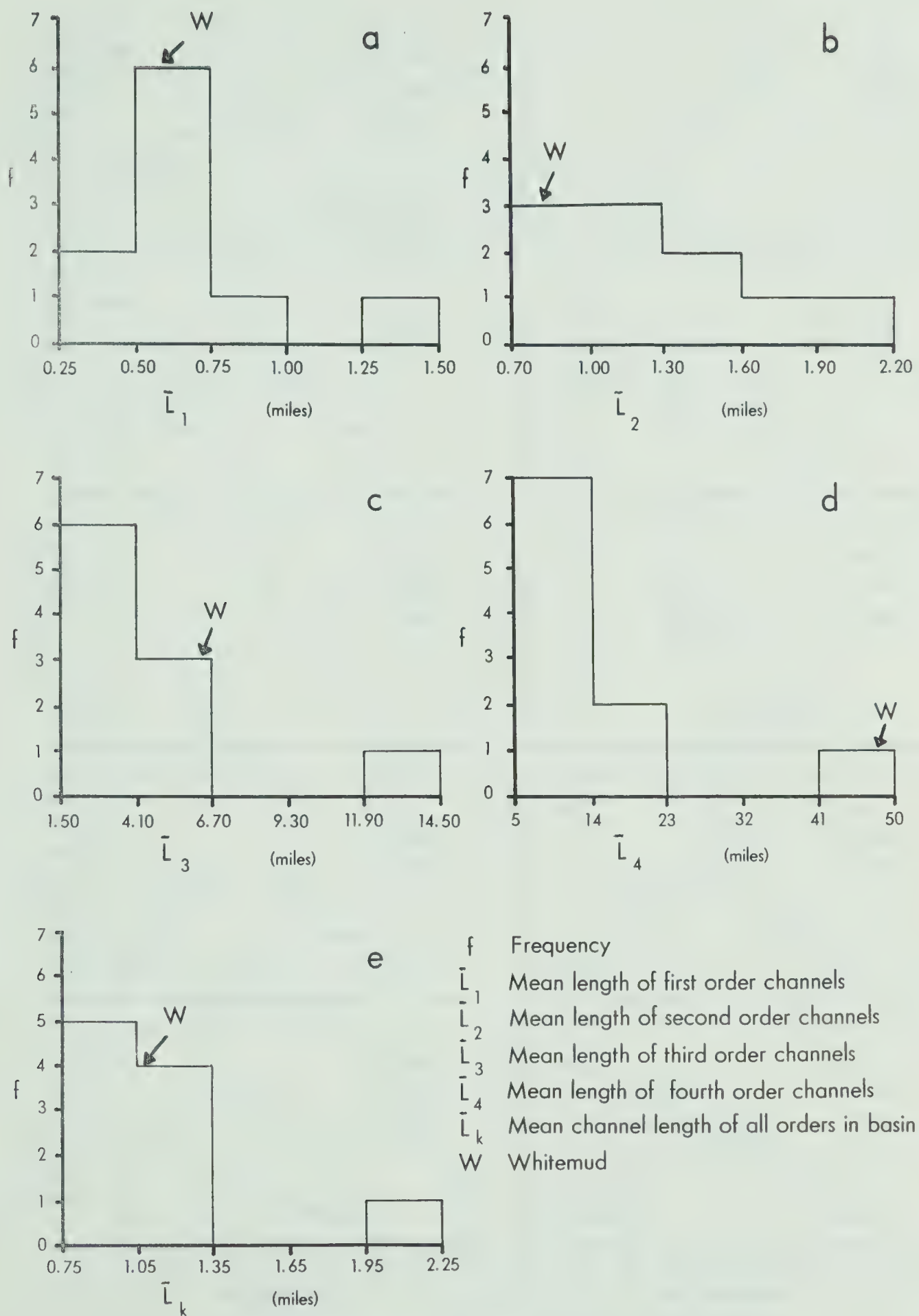


Figure 19 - Mean Channel Lengths

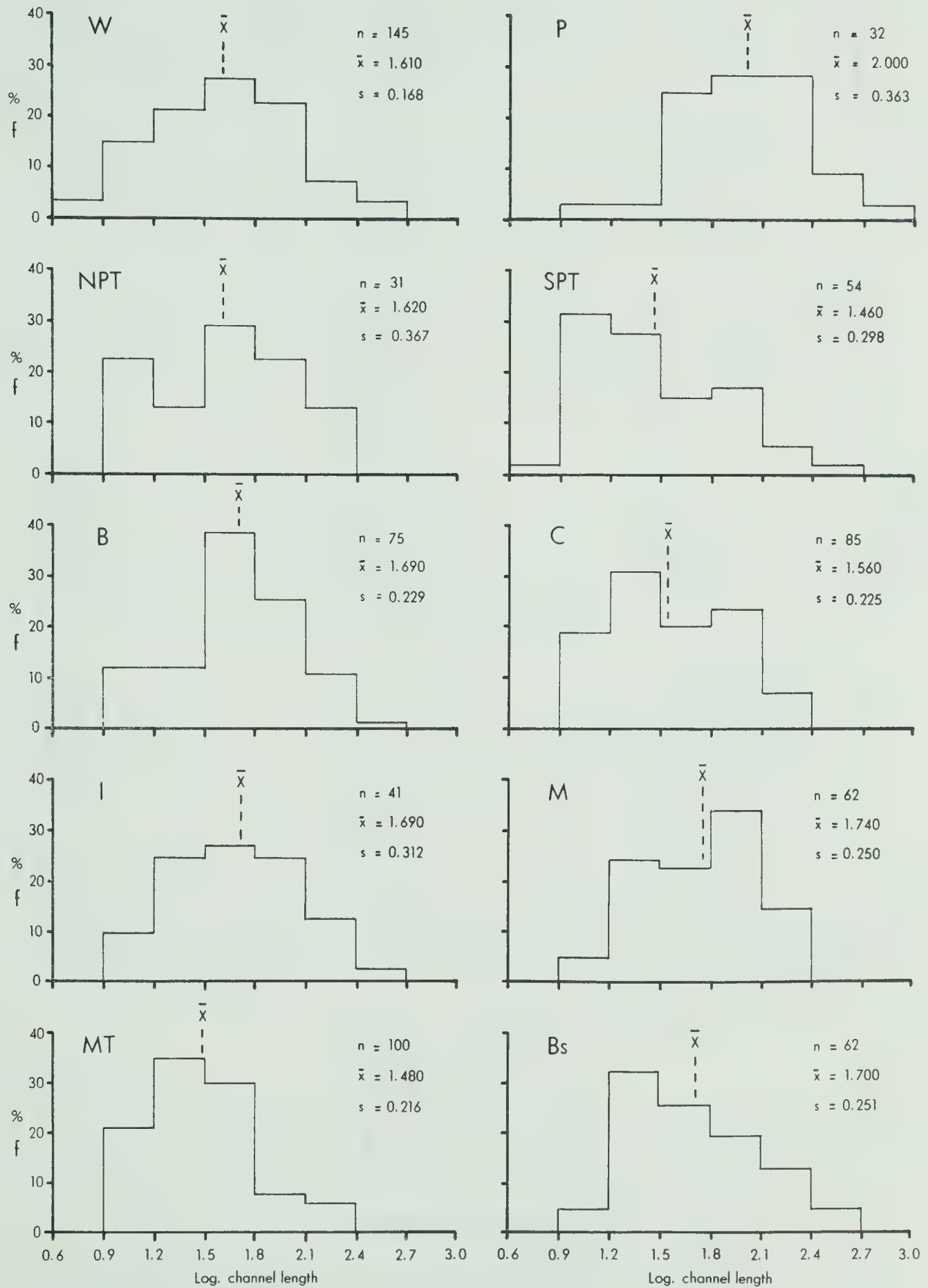


Figure 20 - Logarithmic Frequency Distribution of First Order Channel Lengths

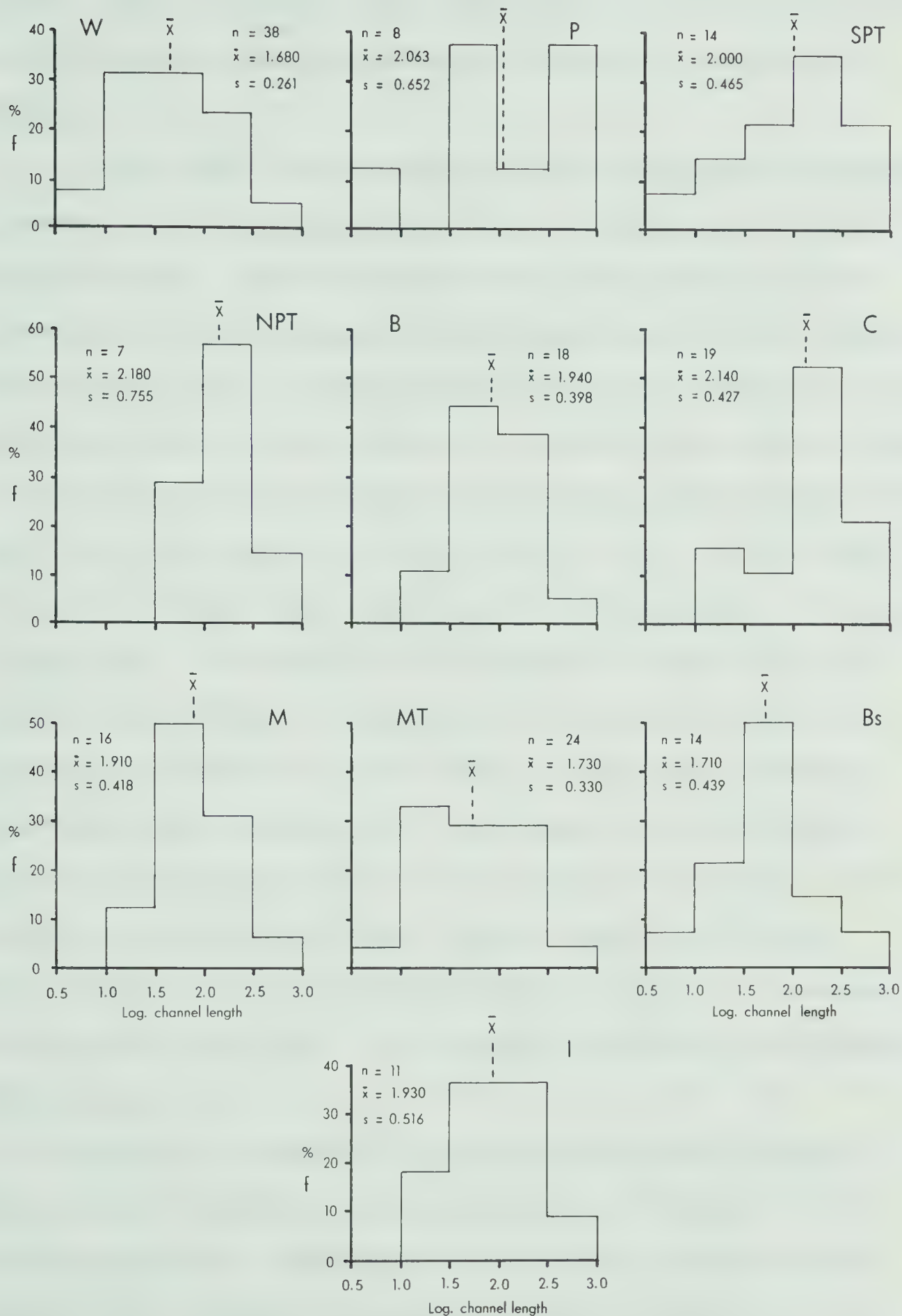


Figure 21 - Logarithmic Frequency Distribution of Second Order Channel Lengths

(Strahler, 1954). Generally, only small numbers of channels are evident for the third order (and, by definition, the fourth) so these were omitted.

A simple visual test of normality of the transformed distributions was carried out by the conventional method of plotting on arithmetic probability paper. The greatest deviations from straight line plots occurred for the South Pipestone Tributary and Bigstone basins (first order channels) and the Pipestone basin (second order channels). Because total populations of channel lengths for each basin were employed such a comparative technique as analysis of variance was not applicable here. It may merely be stated, from Figures 20 and 21, that the logarithmic mean lengths of first order channels appear to be remarkably consistent for most basins. An exception is the Pipestone catchment which has relatively long channels of that order. In this instance, then, the logarithmic transformation did not alter the conclusion, prompted by the arithmetic distributions, that the Whitemud first order channel lengths are quite representative of the study area. Similarly, it may be noted from the logarithmic mean values of second order channel lengths (Figure 21) that the Whitemud basin has second order channels slightly shorter than usual. This also was indicated by the arithmetic distributions.

It may be concluded that for comparative purposes arithmetic means of channel lengths are useful, regardless of skewness of the variate distributions. However, the value of the logarithmic transformation lies in giving a comparative measure of dispersion — the standard deviation. Standard deviations of (logarithm) first order

channel lengths range from 0.168 for the Whitemud basin to 0.367 for the North Pipestone Tributary catchment. Those of the second order range from 0.261 for the Whitemud to 0.755 for the North Pipestone Tributary (Figure 21). Comparisons of standard deviations for distributions with differing means may be expressed by the coefficient of variation (V), or simply the standard deviation divided by the mean, usually converted to percentages (Croxtan and Cowden, 1955). Values of V, describing the relative dispersion of each set, are contained in Table VI.

TABLE VI - COEFFICIENTS OF VARIATION OF MEAN CHANNEL LENGTHS

Basin	V (%) First Order Channels	V (%) Second Order Channels
Whitemud	10	16
Pipestone	18	32
N.P.T.	23	35
S.P.T.	20	23
Blackmud	14	21
Clearwater	14	20
Irvine	19	27
Maskwa	14	22
M.T.	15	19
Bigstone	15	26

The relative dispersion values support the absolute dispersion figures in indicating that the Whitemud first and second order channels deviate least from their mean length, while those of the North Pipestone Tributary show the greatest aberrations. This in part reflects differences in the size of each basin population (N). In local fourth order catchments the logarithmic range and mean length of first and second order channels tends to be reasonably constant regardless of the

number of channels of those orders in each set. From this it follows that the standard deviation decreases with increasing values of N . While the maximum channel segment lengths for each order have been included in Table V out of interest, it is considered that because these largely reflect localized conditions only they are of limited comparative use. Therefore, no further comment need be made in their regard.

Length ratios (Table V) for the central Alberta basins have a grand mean of 3.01, in close agreement with the statement of Strahler (1969, p. 486) that "the mean length of stream segments, in miles, increases by a ratio of roughly three times with each increase in stream order." It may be reiterated that the Whitemud basin has an exceptionally long fourth order channel and third order channels here tend to be a little longer than most of the study basins. As a result, the Whitemud length ratio is of considerably greater magnitude than the others calculated.

Gross (total) channel sinuosity data for the various channel orders ($\bar{G}_1 \dots \bar{G}_4$) in each of the ten basins are contained in Table VII, Figure 22 and Figure 23. Average values of sinuosity for each basin ($\bar{G}_k$) are shown in Table VII and Figure 23.

A comment regarding the use of average ratios should be made at this point. Measurement of the straight-line channel segments, as defined, has the effect of reducing frequency distribution skewness, although not completely eliminating it. It has been shown previously that, for the local basins at least, arithmetic mean values of channel length do have comparative utility notwithstanding

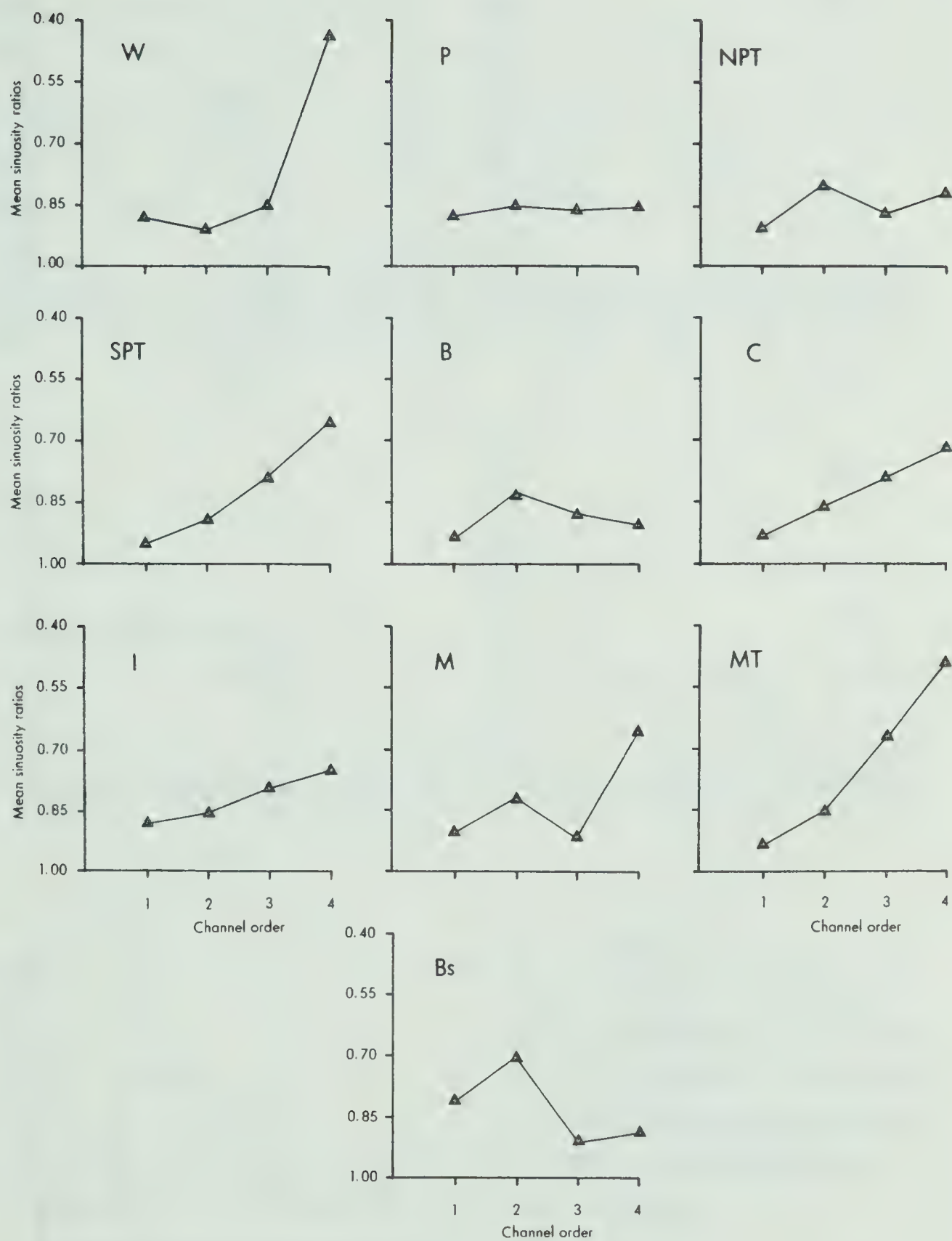


Figure 22 - Mean Sinuosity Ratios for Each Order in the Study Basins

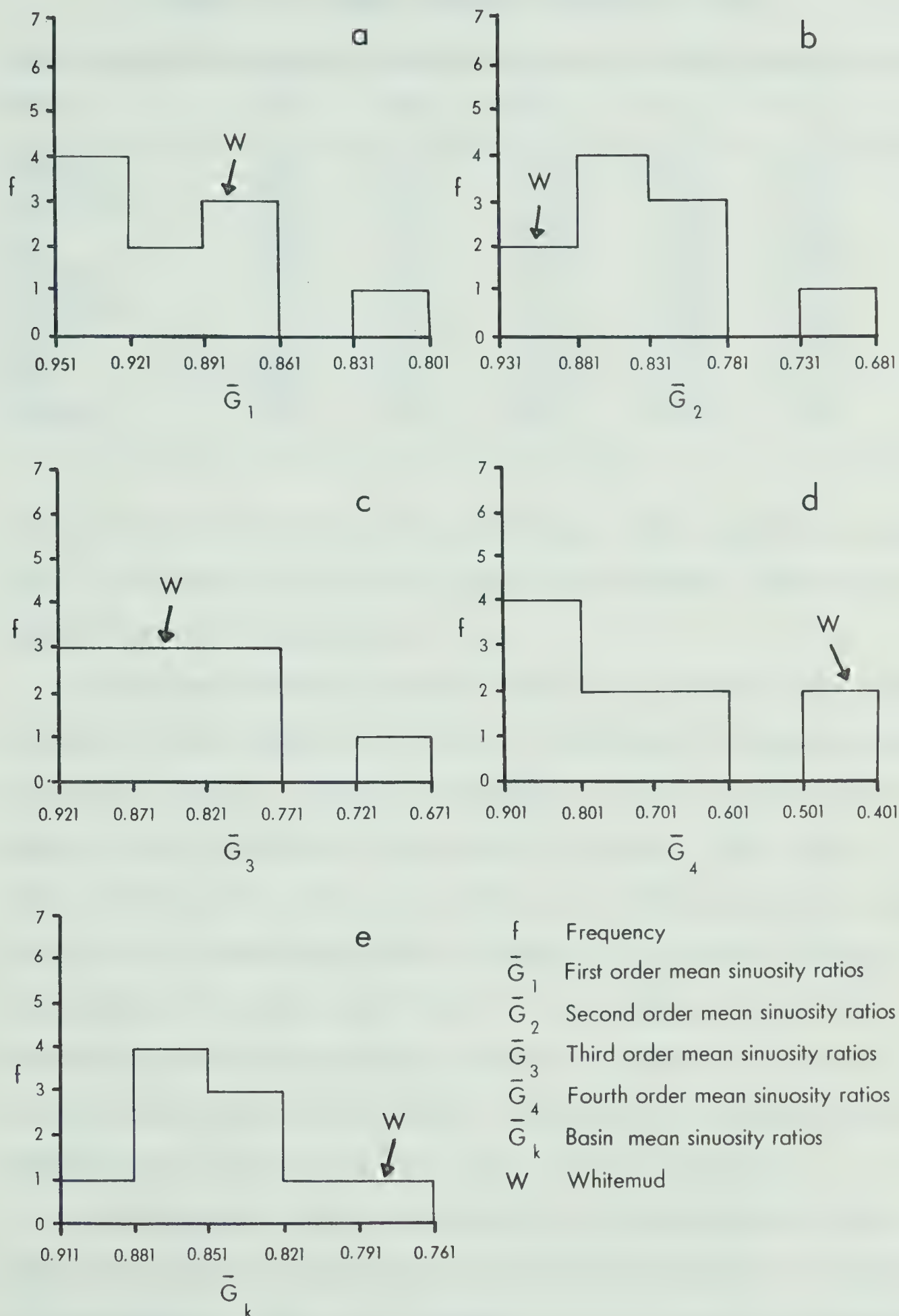


Figure 23 - Mean Sinuosity Ratios

TABLE VII - CHANNEL AND BASIN SINUOSITY RATIOS

Basin	$\bar{G}_1$	$\bar{G}_2$	$\bar{G}_3$	$\bar{G}_4$	$\bar{G}_k$
Whitemud	0.88	0.91	0.85	0.44	0.78
Pipestone	0.87	0.84	0.86	0.85	0.86
N.P.T.	0.90	0.80	0.87	0.82	0.86
S.P.T.	0.95	0.89	0.78	0.65	0.85
Blackmud	0.93	0.83	0.88	0.90	0.90
Clearwater	0.93	0.86	0.79	0.72	0.86
Irvine	0.88	0.86	0.80	0.75	0.84
Maskwa	0.90	0.82	0.92	0.66	0.84
M.T.	0.93	0.86	0.68	0.49	0.81
Bigstone	0.81	0.70	0.91	0.89	0.86

the log-normal distributions of the variates. With a reduction of skewness in the case of straight-line lengths the arithmetic means may be used with some confidence then.

As channel sinuosity decreases the ratio of straight-line length to natural channel length will increase. Therefore, for a given scale of base map, channels exhibiting no sinuosity will be represented by a ratio of 1.00. The effect of map scale is critical. Few, if any, natural channels are perfectly straight and, accordingly, a ratio value of unity may be deemed meaningless. However, provided that the map scale employed is clearly specified (in this instance 1:50,000) the measure has meaning because small elements of sinuosity will tend to be obliterated by map generalization. On this basis, comparisons between different systems, at a given scale, may be carried out.

Theoretically, the largest channels (or highest order) should exhibit the greatest sinuosity because hydraulic sinuosity development will have enhanced that portion bestowed by initial terrain irregulari-

ties. Conversely, first order channels should be relatively straight; a function of both short length (with decreased probability of deviation) and small degree of hydraulically induced sinuosity. Whether an increase in sinuosity with order should approximate a linear, exponential, or some other function is not clear to the writer, although at least half of the study basins tend to have somewhat linear trends (Figure 22). Total sinuosity of first order channels in these basins, with the exception of the Bigstone catchment, has only a small dispersion but considerable scatter exists for other orders (Table VII). In the context of individual basin sets the greatest deviations from a systematic increase of sinuosity with order are characteristic of the North Pipestone Tributary, Blackmud, Maskwa and Bigstone catchments. It would appear from the trend of coordinates (Figure 22) that the second order channels of the North Pipestone Tributary are for some reason more sinuous than might be expected. Reference to Figures 2 and 8 will show that these second order channels act mainly as links between lakes in an area of low-relief, lacustrine deposits. Initial topographic sinuosity has, therefore, played a very dominant role in the formation of these channels, producing very marked channel sinuosity. Again, second order channels of the Blackmud basin seem excessively sinuous, while the main one is relatively straight. Here also the second order channels, in the main, connect small lakes located in low-relief, undulating till zones (Figures 2 and 10). Conversely, the main channel, which is confined to the large Gwynne Channel (Bayrock and Hughes, 1962), had the major portion of its length generalized by straight-line linkage of most probable bifurcation points within the limits of Saunders and

Ord lakes (Section 1:5). This satisfactorily explains sinuosity discrepancies evident in the Blackmud basin. Third order channels in the Maskwa catchment are in general relatively short (Figure 18) and this probably accounts for their low degree of sinuosity. It is not easy to explain why the third and fourth order channels of the Bigstone basin are less sinuous than those of the two lower orders. The majority of the latter are, however, located in moraine covered, bedrock terrain of relatively high relief. Channel distribution is rather chaotic here because of the irregular terrain and this may be the main contributor to high sinuosity of low order channels.

The up-concavity of the Whitemud coordinates in Figure 22 seems closely related to the very long fourth order channel and slightly short second order channels in this basin. Further, the main Whitemud channel has the most advanced level of meander development of any of the basins studied. Figure 23 indicates that the Whitemud basin first order channels are slightly more sinuous than in many of the catchments but a modal class is only poorly developed and the difference may not be significant. The relatively low sinuosity of second order channels in this basin may well be a function of their short length. Third order channel sinuosity is quite typical.

It may be noted from Table VII and Figure 23 that average channel sinuosity ratios for entire fourth order basins have a pronounced tendency to fall between 0.84 and 0.86. Only three of the ten study basins have average sinuosity ratios outside this range. Of these three the Whitemud and Maskwa Tributary basins have generally more sinuous channels and the Blackmud catchment has comparatively

straight stream courses. These differences are largely explained by the variation in sinuosity of the third and fourth order channels in the aberrant basins.

2:4:4 Summary

The preceding analysis of channel length and sinuosity data has shown that the Whitemud basin is atypical in comparatively few departments. Both the length and sinuosity of the fourth order channel are substantially greater than their counterparts in other basins. Third order channels are also relatively long but their sinuosity may be considered typical of all channels of that order in the study area. Average channel lengths of the first and second orders of this basin are not unusual, nor are their sinuosities particularly aberrant. Due largely to the very sinuous main channel the Whitemud basin average sinuosity ratio is larger than usual.

The persuasive theoretical arguments (Section 2:3:2) presented by Shreve (1966) and Milton (1966) may well have great substance but it would seem that the regularities exhibited in average lengths of channels reflect some control(s) of channel development by physical factors. The question here is, are the pronounced modal classes of Figure 19 a result merely of statistical probabilities or are these in part determined by interaction of elements of soil, geology, climate and vegetation? That no simple answer exists is obvious from conflicting viewpoints evident in the literature. To the writer at least the concept of purely random channel network development is not as intuitively attractive as the idea of partially controlled development. Leaving aside the so-called laws of stream numbers and stream lengths, which

may in fact be statistical rather than physical entities, it seems that the regularity of average channel lengths of first order channels (unbranched, and therefore not intimately related to preceding order definition as are say, fourth order segments) in this area reflects some degree of developmental similitude within a reasonably homogeneous physical environment.

2:5 Area and Area/Channel Length Properties

2:5:1 Introduction

Measurement of basin areas may be achieved with a planimeter but in cases where base maps are very large it is more convenient to use grid-squares constructed to the same scale as the maps. Although Monkhouse and Wilkinson (1956) considered this a tedious method numbering of certain rows of squares hastens the addition process. The grid-square method also obviates photographic reduction of large maps to be measured, or the division of these into small units susceptible to planimeter measurement. In this instance the grid employed was divided into units of one quarter of a square mile ensuring accurate estimates of area.

Most of the parameters to be presented in this section are derived by calculating ratios of certain channel length properties to basin area. Therefore, no problems of measurement are involved; rather, the reliability of these calculations depends only on the accuracy of initial channel length and basin area measurement.

2:5:2 Property Definitions

The simple property of basin area needs little further comment other than that it has been observed (Miller, 1953; Schumm, 1956) that

basin areas for a given order usually have extremely right-skewed frequency distributions. Stichling and Blackwell (1957) discussed a basic problem in the delimitation of basin area of prairie catchments, that of depressional storage. In this regard Gray (1964, p. 159) stated:

Most large watersheds contain within their topographic divide a large number of small catchments or sloughs. These sloughs, by their reservoir action tend to modulate the peak discharge from their drainage areas and thus reduce the flood peaks on the main drainage basin.

Thus, the total area contributing runoff to the main channel may vary with differing storm conditions, snow accumulation and soil moisture. Small slough catchments may contribute runoff to the main channel under certain circumstances, but at other times may not. The basin areas delimited here are regarded as maximum possible values.

A most important measure is that of drainage density, defined by Horton (1932) as the average length of streams within a basin, per unit of area. In regions of homogeneous lithology, climate, and stage of development, drainage density values tend to remain constant and are independent of basin size (Melton, 1957). Horton (1945) noted that the accuracy of this measure is dependent upon a map scale sufficient to show all permanent natural stream channels. It is, however, not necessary that the channels be permanent, if indeed any are in nature, but only that the method and scale of channel delimitation be unambiguously specified. The inverse of drainage density was defined by Schumm (1956) as the constant of channel maintenance, or the number of square feet (or square miles) of basin surface needed to maintain one foot (or mile) of channel.

A further measure of drainage network texture is that of channel frequency. Horton (1932) stated this to be the total number of channels in a basin divided by the basin area. Channel frequency values will differ for a given network according to whether the Horton or Strahler ordering system is used. Allied to this measure are the properties known as the frequency of first order channels and density of first order channels (Melton, 1957). These are simply defined in terms of total number and total length of first order channels divided by the area of the higher order basins being studied. Melton (1957) considered that these could be used together to determine whether increases in drainage density result from headward growth of channels or from an increase in first order channel numbers.

The peripheral area of a basin, not reached by first order channel encroachment, is of considerable interest (Section 2:4:2). In very general terms this area might be thought of as akin to Horton's (1945) "belt of no erosion." An estimation of the channel-penetrated area may be obtained by taking the following steps. First, generalize the channel network by joining all channel heads and bifurcation nodes by straight lines. Second, working around the outer limit of the channel network join consecutive first order channel heads by straight lines. After delimitation of the penetrated area this may be measured and calculated as a percentage of the total basin area. Although not entirely free of inconsistencies (discussed in Section 2:5:3) this method is useful in furnishing approximate comparative values, and it may serve to distinguish unusual basins in a given region.

2:5:3 Central Alberta Data

Appearing in Table VIII are values of basin area (A), drainage density (D), constant of channel maintenance (C), channel frequency (F), frequency of first order channels (F_1), density of first order channels (D_1) and the percentage of A into which channels have penetrated ($A_p\%$).

TABLE VIII - AREA AND AREA/CHANNEL LENGTH PROPERTIES

Basin	A (sq.mi.)	D (miles per sq.mi.)	C (sq.mi. per mi.)	F (no.per sq.mi.)	F_1 (no.per sq.mi.)	D_1 (miles per sq.mi.)	$A_p\%$
Whitemud	150.25	1.35	0.74	1.26	0.97	0.59	51
Pipestone	131.25	0.73	1.37	0.33	0.24	0.35	37
N.P.T.	58.00	0.87	1.15	0.77	0.58	0.37	57
S.P.T.	84.25	0.98	1.02	0.85	0.64	0.31	47
Blackmud	61.75	1.34	0.75	1.57	1.21	0.80	50
Clearwater	100.75	1.08	0.93	1.08	0.84	0.46	45
Irvine	59.00	0.99	1.01	0.95	0.69	0.45	58
Maskwa	64.00	1.30	0.77	1.28	0.97	0.73	42
M.T.	55.00	1.79	0.56	2.36	1.82	0.73	44
Bigstone	68.75	1.31	0.76	1.16	0.90	0.74	54

Areal values for the ten local basins exhibit the marked right-skewness noted by previous investigators and the frequency distribution has pronounced monomodality (Figure 24a). While local fourth order basins tend to have areas between approximately 50 and 75 square miles, the Whitemud basin is particularly atypical in this regard, with an area more than twice the modal value. However, the importance of area differences for basins of a given order should not be unduly stressed because drainage basin boundaries in an area dominated by relatively young glacial and glacio-lacustrine deposits were determined in part

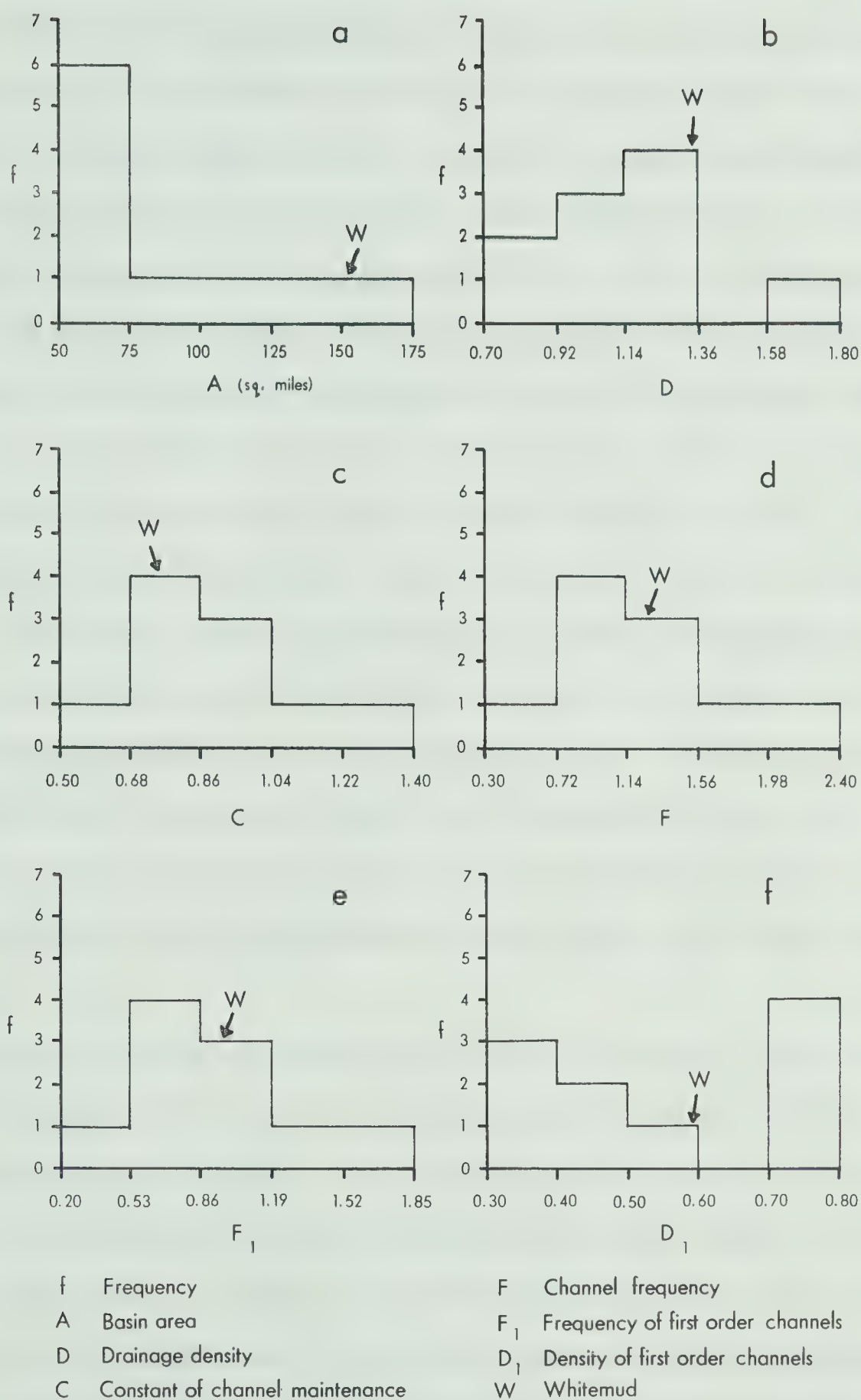


Figure 24 - Area and Area/Channel Length Properties

by the spatial arrangement of depositional "highs" and "lows." The basic outline form of medium order drainage basins in this area is a direct result of the interaction of glacial deposition onto a pre-glacial, fluvially eroded surface. Lacustrine deposition during stages of deglaciation further modified the terrain. Variability in till deposition especially may have determined the location of the greater portion of the divide zones. While minor changes in the position of these may have resulted from subsequent operation of slope and channel processes, in general the drainage networks are as yet too far removed from divides to have exerted much influence on their location. Second, the concept of basin order as a fundamental comparative base cannot be wholly consistent. Shreve (1966) pointed out that the strategic addition of one first order channel in a basin of, say, fourth order may be enough to increase the order of the entire basin by one. This does not negate the usefulness of the order concept but it emphasizes that inter-regional or inter-basin comparisons should be determined mainly by measures which are largely independent of basin order, for example drainage density.

Before further consideration of the local drainage densities it is instructive to look at values calculated by others for differing North American environments. In his comprehensive review of morphometric methodology and published work Strahler (1964) noted that in the U.S.A. low values of D tend to be in the range of 3.00 to 4.00, for example, in the Appalachian Plateau area where resistant sandstones occur. He considered that in the humid central and eastern U.S.A. values of D range between 8.00 and 16.00 where moderately resistant

rocks and a deciduous forest cover are present. Farther west, in the Rocky Mountains and coast ranges of Southern California this property exhibits values ranging from 8.00 to 100.00, varying mainly with lithology, climate and vegetation differences. Extreme drainage densities, sometimes over 1000.00, have been found in badland environments (Schumm, 1956).

Accepting the fact that the scale on which channels have been recognized is smaller than in most previous work, it is obvious that local drainage densities are particularly low and are indicative of very coarse-textured erosional terrain. Strahler (1964) stated that "low drainage density is favored in regions of highly resistant or highly permeable sub-soil materials, under dense vegetation cover and where relief is low." While the Cretaceous and Tertiary bedrock and the Pleistocene sediments cannot be considered as either highly resistant or highly permeable, the vegetation and relief elements favorable for low values do occur here. Perhaps of greater significance, however, are the climatic factors which produce minimal runoff. Melton (1957) has shown that drainage density is very strongly correlated with Thornthwaite's (1931) precipitation-effectiveness index. This index, which is dependent on the amount and distribution of precipitation and evaporation, provides a measure of the availability of moisture to vegetation (Melton, 1957, 1958a, 1958b). Therefore, although the lithologic character here is of a relatively non-resistant variety, the interaction of low relief, a relatively dense natural vegetation (grass or boreal forest, or transitional combinations of these) and especially a climate producing little surface water for runoff (except

for limited periods each year) has maintained an extremely coarse-textured erosional landscape.

Melton (1958b, p. 35) has stated that:

within an area of homogeneous lithology, climate and erosional history, values of D and F vary closely about a mean value for portions of the area that are large with respect to the size of the average first order drainage basin. Generally, D and F show no systematic variation with size of area, above the maximum size mentioned. That is, topographic texture is a property of the region rather than of individual basins in the region.

This contention would seem to be borne out within the central Alberta study area where drainage densities are quite stable from one basin to the next; minor exceptions are the Pipestone and North Pipestone Tributary basins with slightly lower values and the Maskwa Tributary catchment having a drainage density higher than usual (Table VIII and Figure 24b). The Whitemud value falls in the modal class which is not well-pronounced. Indeed, the drainage densities appear subjectively to approach a normal distribution, as may be true for a region of homogeneous lithology (Melton, 1957). On the other hand, Maxwell (1961) concluded from his data that drainage densities are log-normally distributed.

The frequency distribution of C values appears to be somewhat right-skewed (Figure 24c). Again, the Whitemud basin cannot be considered atypical in this context. It may be noted from Table VIII that the most common areas of basin surface needed to maintain one mile of channel lie between three quarters to one square mile. Not unexpectedly, because C is the reciprocal of D, the Pipestone, North Pipestone Tributary and Maskwa Tributary basins deviate most from the modal class.

While drainage density and channel frequency are both measures of terrain texture they differ and in theory one may be largely independent of the other. This may be particularly true of areas which might be termed "youthfully developed" having many swamps, sloughs, and poorly-drained zones (Melton, 1958b). However, Melton (1958b) has shown that for well-developed ("mature")⁹ drainage basins F and D are related in terms of a power function expressed by the equation $F = 0.694 D^2$. Therefore, geometrically similar basins should have similar values of $\frac{F}{D^2}$. It is surprising that at least three of the local basins yield $\frac{F}{D^2}$ values fairly close to the observed constant of 0.694, the Whitemud value being extremely coincident (Table IX).

TABLE IX - VALUES OF $\frac{F}{D^2}$

Basin	$\frac{F}{D^2}$	Basin	$\frac{F}{D^2}$
Whitemud	0.691	Clearwater	0.926
Pipestone	0.619	Irvine	0.969
N.P.T.	1.017	Maskwa	0.757
S.P.T.	0.885	M.T.	0.737
Blackmud	0.874	Bigstone	0.676

Whether this indicates that the Whitemud has a better-developed channel network than the other basins is conjectural. If "maturity" of basin erosional form is a prerequisite to the validity of Melton's (1958b)

⁹Melton (1958b) considered "mature" basins to be only those in which every channel has developed a watershed with smooth slopes extending to the divides.

observation then the local values closely approximating the observed constant may well occur by chance. In no way may the local basins be considered "maturely" developed in the sense envisaged by Melton (1958b).

Figure 24d shows that in the Whitemud basin stream frequency is a little higher than the modal class but is not sufficiently far removed to be considered as very atypical of the study area. The same is true for the frequency distribution of F_1 values (Figure 24e). The frequency distributions of C , F and F_1 are identical in this case, suggesting some degree of interdependence between these properties, notwithstanding the "youthful" nature of the erosional terrain. However, the density of first order channels is another matter. A most peculiar distribution is apparent (Figure 24f) for this property; a distribution suggesting that perhaps two families of D_1 exist. In any case, the Whitemud value falls between the bi-modes and can, then, be considered somewhat atypical.

Earlier a method of estimating the percentage of basin area penetrated by the channel network was outlined. It was noted that certain problems were encountered in the joining of first order channel heads by straight lines. These may be summarized thus:

- (1) where first order channel heads are located in the interior of the drainage system
 - (2) where for a considerable length a higher order channel is not joined by first order direct entrants on the bank nearest the main divide
- and (3) where part of a straight line joining consecutive first order channel heads falls outside the main divide.

Included as an example is a generalized form of the Whitemud channel network and delimited area A_p (Figure 25). Examples of problems (1) and (2), and means of overcoming them, are illustrated in the areas marked A and B respectively, on the map. In the case of problem (3) only that area included by the main divide and portions of the straight line cutting the divide was incorporated.

Calculation of the $A_p\%$ figure for each basin produced values ranging from 37 per cent (Pipestone basin) to 58 per cent (Irvine basin) with an almost even distribution between the extremes. In the absence of a well-defined modal class all that can be stated is that the Whitemud value of 51 per cent is slightly larger than the regional average of 49 per cent, a difference which is probably not significant because of the technique involved. On the basis of this approximation, though, it seems reasonable to suggest that about half of the surface area of basins studied remain as yet inviolate by channel development.

2:5:4 Summary

In terms of most area/channel length properties the Whitemud basin may be deemed typical of the local region. While the absolute area of this fourth order basin is substantially greater than for most basins of like order it has been noted that the dispersion of absolute areas within a given order is large. The only other major departure from a local norm is evident for the Whitemud D_1 value. However, it is difficult to decide in this instance what should be considered as typical for the area because of the peculiar frequency distribution of the variates.

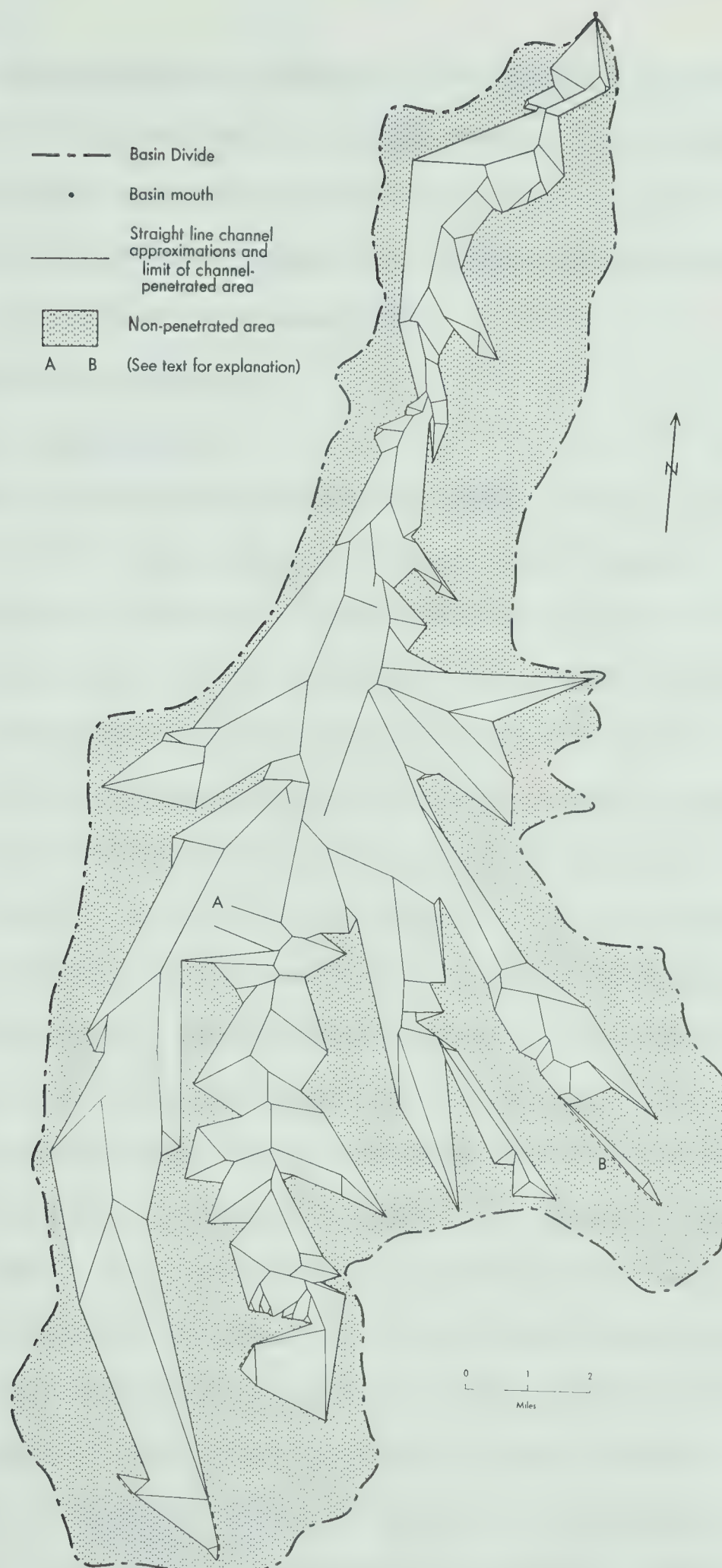


Figure 25 - Whitemud Basin Example of Method of Delimiting Channel-penetrated Area

The new morphometric measure ($A_p\%$) introduced in this study remains to be refined and tested for other areas. It is hoped that the method will prove useful for inter-basin and inter-regional comparisons. In the local context the Whitemud basin appears to be of average stature insofar as this property is concerned.

2:6 Basin Shape Properties

2:6:1 Introduction

Bunge (1962) and Boyce and Clark (1964) strongly advocated rigorous application of shape measures in geographical research. The latter authors noted that theoretical formulations have evolved from analyses of shape, particularly in the biological sciences and in mineralogy. However, in geography shape has been employed primarily as a descriptive device. Geomorphologists and hydrologists have developed various measures in describing and analyzing drainage-basin shape for two basic reasons. First, the geomorphologist is, by definition, interested essentially in form and studies of basin shape may assist his interpretation of past geomorphological events or aid theoretical hypothesizing. Second, the hydrologist has an interest in basin shape because it has been shown in many instances that this factor may be intimately related to various hydrologic characteristics (Gray, 1964; Strahler, 1964). Thus, in establishing measures of drainage basin shape the student of landforms may, in addition to answering questions regarding basin evolution, assist hydrological studies of that area.

A number of basin shape measures have been presented in the literature, all having certain advantages and disadvantages. Since no one measure is yet generally accepted as a standard it was thought

expedient to include most of the more commonly used ones in the present study. Following this, some estimation of the relative merits of each measure could be formulated.

2:6:2 Property Definitions

Horton (1932) proposed a measure of basin shape which he termed a form factor. This was defined as the ratio of basin width to basin length but he stipulated that the basin length should not necessarily be the maximum length. Rather, this was to be taken as the length from the basin mouth to a point on the divide opposite the head of the main stream. Horton (1932) expressed his form factor as being equal to the basin area divided by the square of basin length (as defined above) and stated further (Horton, 1932, p. 351):

this factor has been considerably used in connection with maximum flood-discharge formulas. In the case of long, narrow drainage-basins such, for example, as basins occupying synclinal valleys and rift valleys, the form factor is indicative of the flood-regimen of the stream. For drainage-basins of irregular form, especially those with permeable soils, [the] form factor is not a sensitive indicator of hydrologic characteristics.

Both Strahler (1964) and Morisawa (1968) noted that the United States Corps of Engineers adopted the inverse of Horton's form factor for application to unit-hydrograph data.

Miller (1953) introduced a measure of basin shape which he called the circularity ratio, or the ratio of basin area to the area of a circle with a perimeter length identical to that of the basin.¹⁰ A

¹⁰Basin perimeter may usually be measured accurately with a chartometer because perimeter sinuosity tends to be less accentuated than, for example, certain channels which render chartometer measurement inaccurate (Section 2:4:1).

closely related measure was proposed by Schumm (1956) who defined his elongation ratio as the ratio between the diameter of a circle, with the same area as the basin, and the maximum length of the basin (longest dimension of the basin parallel to the main channel). An element of subjectivity enters in demarcating the basin length as defined by Schumm (1956) and Maxwell (1961) sought to eliminate this by making certain modifications. He re-defined the elongation ratio as the ratio of the diameter of a circle, having the same area as the basin, to the "diameter of the basin." To this end he advocated application of the following criteria for delimiting basin diameter (Maxwell, 1961, pp. 36-37):

Criteria. The diameter must be a straight line and (1) be essentially parallel to the longest drainage line, (2) divide the main channel into segments such that the sums of segment lengths on opposite sides of the diameter are approximately equal, (3) be parallel to the line which separates opposite-facing valley slopes, (4) bisect the basin area, (5) be the longest diameter.

Application. The first criterion listed above is to have the greatest weight and the last the least weight. If the channel is essentially:
straight and longer than one-half of diameter,
 use criteria 1, 2 and 3,
straight and shorter than one-half of diameter
 use criteria 1, 3 and 4,
curved and longer than one-half of diameter
 use criteria 5, 3 and 2,
curved and shorter than one-half of diameter
 use criteria 3, 4 and 5

Criticism has been levelled at shape measures based on circles. Maxwell (1961) commented that the circularity ratio does not discriminate between deviations from perfect circularity resulting from either irregularities in gross basin outline or minor crenulations of the

perimeter. A second criticism is that few drainage basins approach a circular form in nature, being more often somewhat pear-shaped. Chorley, Malm, and Pogorzelski (1957) proposed, therefore, that a lemniscate loop be recognized as a standard for basin shape estimation. Avoiding the somewhat complicated mathematical arguments related to this measure it may be stated that the method yields two measures of shape. The first was defined as the value, k , describing a relationship between maximum basin length and maximum basin width. This may be calculated according to the following expression (Chorley, Malm, and Pogorzelski, 1957, p. 139):

$$k = \frac{\ell_{\max}^2 \pi}{4 A}$$

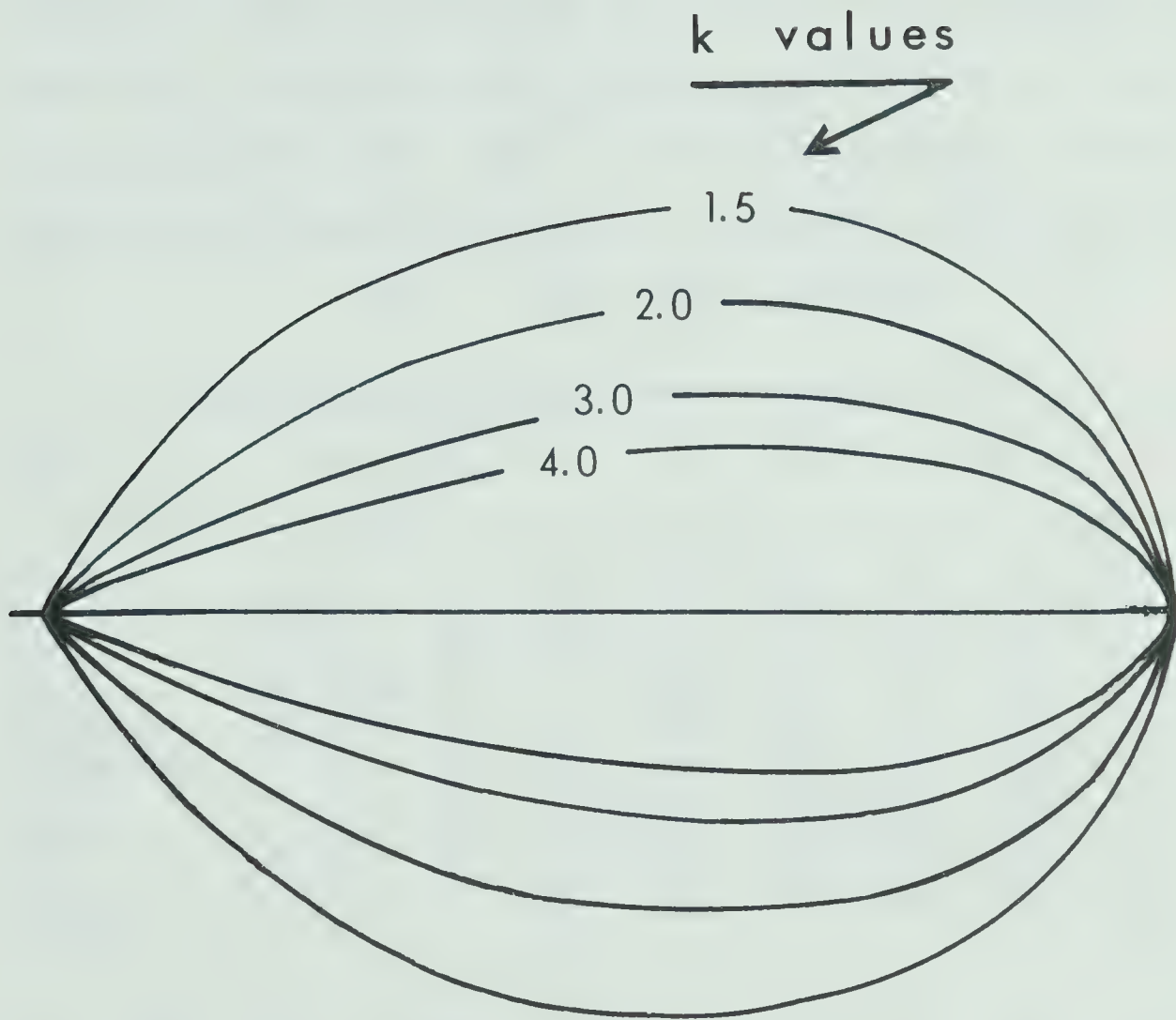
where: $\ell_{\max}$ equals the maximum length of the basin from the mouth to some point on the divide,

π equals 3.1416 and,

A is equal to basin area.

The value, k , describes the ideal lemniscate loop shape most nearly approached by the basin shape. Figure 26 illustrates various lemniscate loops and their related k values. Chorley, Malm, and Pogorzelski (1957) considered that the value, k , might be especially useful for estimating regional tendencies and making inter-regional comparisons.

The second measure takes this analysis one step further in indicating how closely the natural basin shape approximates the related ideal lemniscate loop. This measure, termed the lemniscate ratio by the above authors, embraces a rather tortuous set of procedures involving the use of elliptic integral and arc sine tables. It allows calculation of the ideal lemniscate loop perimeter length which may



Source: Chorley, Malm, and Pogorzelski (1957)

Figure 26 - Ideal Lemniscate Loops

then be compared with the basin perimeter to produce the lemniscate ratio.

2:6:3 Central Alberta Data

Apart from their use in some of the shape measures previously defined the simple properties of basin perimeter (P), maximum basin length ($\ell_{\max}$), and maximum basin width ($w_{\max}$)¹¹ indicate size and shape in a very general sense. Table X contains these values for the local study basins, in addition to values calculated for the Horton (1932)

TABLE X - BASIN SHAPE PROPERTIES

Basin	(in units of miles)				S_1	S_2	S_3	S_4	k	K
	P	$\ell_{\max}$	$w_{\max}$	L_d						
Whitemud	74	27.4	11.6	27.4	0.200	4.997	0.345	0.502	3.924	0.967
Pipestone	73	25.0	10.8	25.0	0.219	4.573	0.310	0.517	3.740	0.964
N.P.T.	39	15.4	5.6	13.8	0.318	3.142	0.438	0.533	3.514	0.959
S.P.T.	43	14.6	8.4	14.6	0.406	2.461	0.573	0.709	1.987	0.864
Blackmud	39	12.0	6.2	11.8	0.539	1.854	0.510	0.739	1.832	0.838
Clearwater	48	17.2	8.4	17.2	0.341	2.936	0.550	0.658	2.306	0.901
Irvine	50	17.6	7.4	13.4	0.225	4.448	0.297	0.493	4.123	0.970
Maskwa	43	13.4	9.6	13.4	0.694	1.440	0.435	0.674	2.204	0.891
M.T.	41	10.2	8.8	10.2	0.561	1.782	0.411	0.820	1.486	0.740
Bigstone	48	17.0	7.8	17.0	0.253	3.960	0.375	0.550	3.302	0.953

form factor (S_1), the Corps of Engineers form factor (S_2), Miller's (1953) circularity ratio (S_3), and Schumm's (1956) elongation ratio (S_4) as modified by Maxwell (1961) using basin diameter (L_d). This

¹¹In this study $w_{\max}$ was determined as the line of maximum basin width drawn at right angles to the line of maximum basin length.

table also includes the lemniscate (k) values and lemniscate ratios (K)¹² based on the method devised by Chorley, Malm, and Pogorzelski (1957). Figure 27 illustrates the frequency distributions of these variates. The Whitemud basin is comparatively large in terms of basin perimeter, maximum basin length, maximum basin width and basin diameter. Related essentially to the comparatively large area of this basin, these measurements add little to what has already been presented and merely substantiate the fact that in absolute terms the Whitemud basin is atypical of its fourth order counterparts.

However, the dimensionless shape ratios have greater meaning for comparative purposes. It may be noted from Figure 27 that, at least within the framework of most of the shape measures employed, the Whitemud basin is not atypical. It remains to look at each of these measures in detail.

A minor difficulty arises in calculating the Horton form factor if the Horton system of channel ordering is not employed. It becomes necessary to decide where the head of the "main channel" lies so that the basin length as defined by Horton (1932) may be measured. With the Strahler channel ordering system the "main channel" heads are

¹²The reader should note that the lemniscate ratio has been expressed here as K , to facilitate consistency in presentation. In the original work of Chorley, Malm, and Pogorzelski (1957) this letter represented another value in the computation procedure which should not be confused with the K values as defined in the present study.

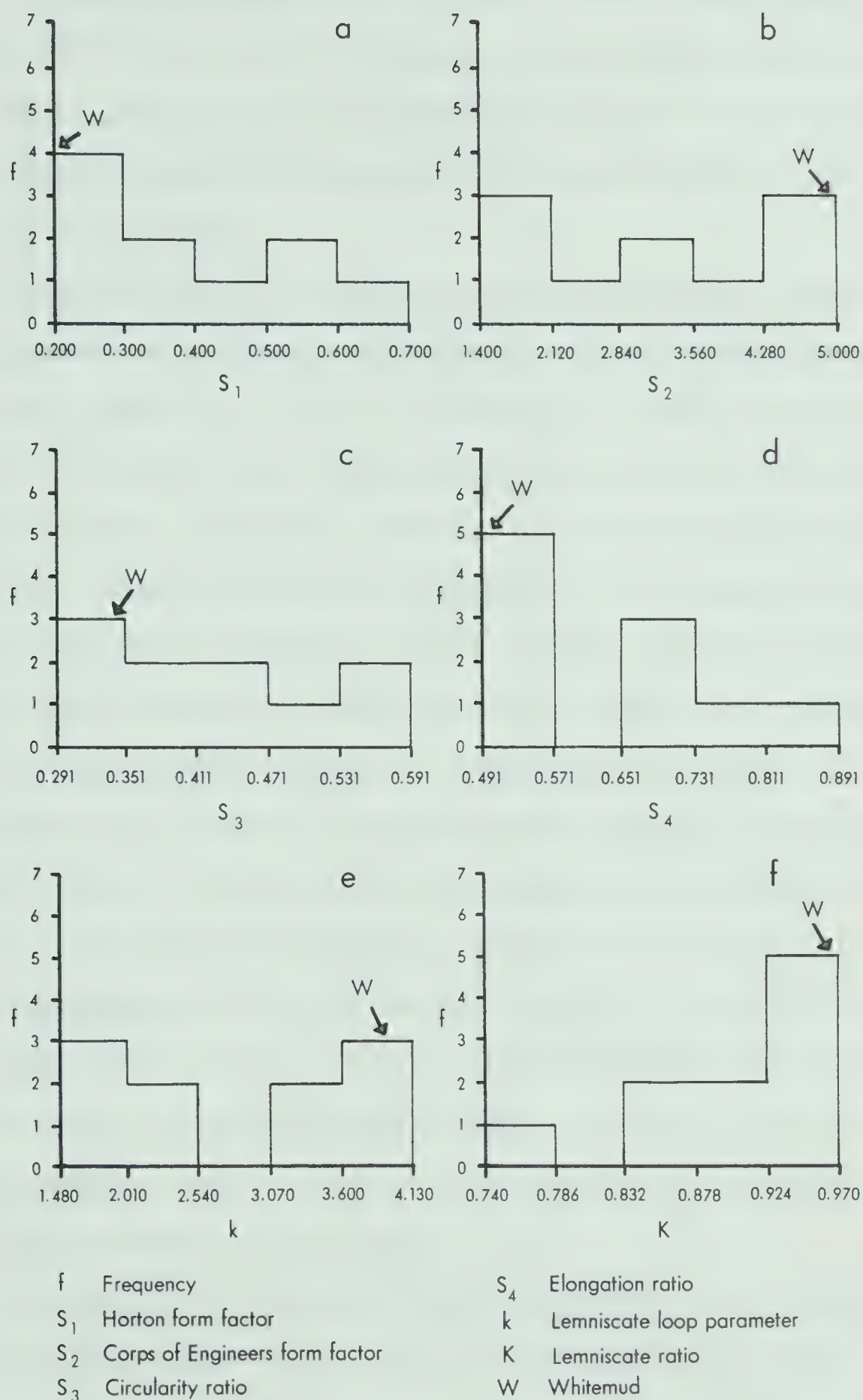


Figure 27 - Basin Shape Properties

usually located toward the basin interiors. To allow consistency with Horton's (1932) form factor definition, however, basin length was measured from a point on the divide opposite the head of the longest flow path. This is equivalent to adopting the Horton ordering system in this particular instance.

Determination of S_1 for the local basins produced a mono-modal, right-skewed distribution with the Whitemud value occurring in the modal class (Figure 27a). The S_1 values may be interpreted thus; an increase in the form factor value stems from diminished basin length relative to width. Therefore, the modal class containing the Whitemud basin value represents relatively long basins (the Whitemud, Pipestone, Irvine and Bigstone catchments). While this ratio appears to be of value in describing and differentiating basin shapes here, the method may not be particularly suitable for future hydrologic work. This is because the local basins are irregular in form and have reasonably permeable soils — two conditions which apparently make the form factor an insensitive indicator of hydrologic variables (Horton, 1932).

Computation of the S_2 values and plotting of these in a histogram (Figure 27b) resulted in a bi-modal distribution, with the Whitemud figure in one of the modal classes. While this might possibly be interpreted as indicative of two shape families the poorly-accentuated bi-modes render this conjectural.

The frequency distribution of S_3 has a rather even spread throughout the range of values. Thus, the weakly-pronounced modal class, containing the Whitemud basin value, may not be confidently designated as the most typical. Both Morisawa (1958) and Maxwell (1961) commented

that the circularity ratio may be significantly increased or decreased by varying degrees of perimeter sinuosity without concomitant, significant changes in basin area. It would appear that variability of local basin perimeter sinuosity may have been instrumental in producing the widely-dispersed distribution evident in Figure 27c. However, it is still interesting to note that the mode, poorly-expressed as it is, indicates the largest single class frequency is that for relatively long basins.¹³ Horton's (1932) form factor demonstrated this as well.

Turning now to the closely related elongation ratio (S_4) we may see (Figure 27d) that a well-defined modal class emerged for the local data. Again the Whitemud figure is included in the modal class and may, therefore, be considered typical. With this measure a value of unity describes a perfect circle (Schumm, 1956). Strahler (1964) was of the opinion that low-relief basins characteristically have elongation ratios close to unity and that where steep ground slopes exist values between 0.60 and 0.80 are common. In the central Alberta study area, which is of generally low relief, elongation ratios are, however, much smaller than might perhaps be expected. This measure, also, corroborates the previous finding that many of the local fourth order basins tend to be relatively elongated.

¹³Strahler (1964), summarizing Miller's (1953) findings, stated that strongly elongated basins in the Clinch Mountains, Virginia, tended to have circularity ratios between 0.40 and 0.50. From this it may be suggested that most of the central Alberta basins studied are strongly to very strongly elongated (Table X and Figure 27c).

Finally, the lemniscate-based measures of basin shape should be commented on. Values of k in this case have a bi-modal distribution (Figure 27e), possibly indicative of a two-family shape component. As for the S_2 shape measure, also exhibiting a bi-modal distribution, it is conjectural whether the Whitemud basin may be considered typical or not in this regard. If, however, the confidence of Chorley, Malm, and Pogorzelski (1957) in k as a measure of regional tendency is not misplaced, then we might assume that fourth order basins here do indeed tend to fall into either of two shape families. It follows, then, that the Whitemud basin is typical of one of these families (Figure 27e), and that this family is composed of the more elongated basins (Figure 26).

The above authors did not seem to be so well-disposed toward the lemniscate ratio as they were to the k value, as a comparative measure. It is obvious that exactly the same criticism may be made of this measure as has been levelled at Miller's (1953) circularity ratio; namely, the values of K may be significantly effected by minor crenulations of the basin perimeter. Figure 27f would appear to show that many of the local basins differ only slightly from their ideal lemniscate counterparts since their K values, including that of the Whitemud basin, are close to unity. A value of unity here indicates perfect coincidence of the basin perimeter length with the k -determined lemniscate loop perimeter. However, perimeter length coincidence does not necessarily mean shape coincidence and thus this measure loses some of its appeal.

Work carried out by Morisawa (1958) in the Appalachian plateau area is of interest. She tested five shape measures, S_2 , S_3 , S_4 and k , as defined here, and another, the ratio of maximum basin length to maximum basin width, for their correlation with runoff from twenty-five basins. The results of her regression analyses led Morisawa (1958) to conclude that both the elongation ratio, and particularly the circularity ratio might be considered of importance for hydrological studies.

Little may be stated as to the relation of the local basin shapes to hydrological characteristics because of the present paucity of streamflow and runoff data. It is hoped, however, that these shape measures might be tested, for at least the Whitemud basin, as soon as sufficient data have accumulated.

Before the conclusion of this discussion of basin shapes it is necessary to consider possible reasons for the existence of predominantly elongated catchments in this area. As stated earlier, in the writer's opinion the local fourth order basin boundaries have not been determined by fluvial erosion processes. Rather, it seems most likely that the main elements of basin outline form have resulted from:

- (1) the position of preglacial divides and other relatively elevated portions of the terrain,
- and (2) the subsequent modification of this preglacial terrain by differential glacial erosion and deposition of glacial and glacio-lacustrine sediments. Mass-waste processes may have created further minor adjustments of the basin boundaries in early postglacial time especially.

Farvolden (1963a) prepared a map of the bedrock topography for a large area of central Alberta, including all of the present study area. This information is shown in Figure 28 and although it is necessarily of a general nature it is of value. Of the ten fourth order basins selected only the North and South Pipestone tributary catchments do not appear to have quite well-defined preglacial valleys or depressions in close proximity to portions of the present basin axes (parallel with the present main channels). The form of preglacial valleys has possibly been the most important single factor determining the outline configuration of present fourth order basins. Farvolden (1963b) considered that dendritic and well-integrated drainage systems were developed here prior to glaciation. Whatever processes tended to produce relatively long basins at that time, their imprint apparently persists in the form of the present catchments.

2:6:4 Summary

The application of various measures of drainage basin shape has indicated that the central Alberta catchments tend to be relatively elongated. That this may have certain effects on the hydrological character of individual basins is very likely but as yet these effects cannot be determined because of a lack of hydrologic data. The possibility that the local basins fall into two distinct shape families cannot be entirely discounted since two of the shape-measure frequency distributions exhibited some degree of bi-modality. It is considered, also, that the shape characteristics here probably stem not from fluvial erosion processes but have been, in the main, determined by the preglacial terrain.

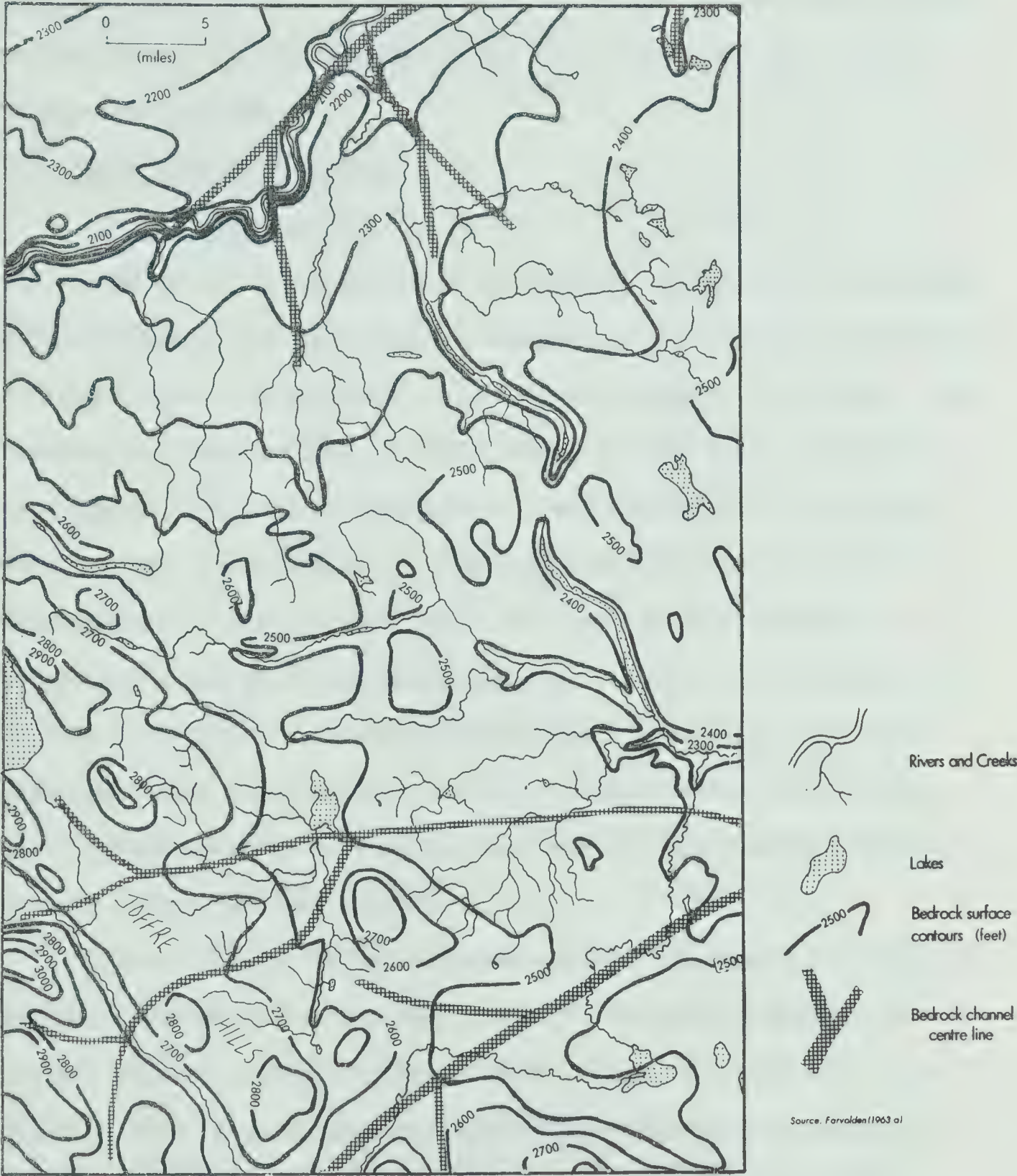


Figure 28 - Bedrock Configuration of the Study Area

The Whitemud basin may be considered typical in terms of dimensionless shape ratios, if not in absolute dimensions of length, width, and perimeter. In the context of the latter measures this basin is comparatively large.

2:7 Basin Relief Properties

2:7:1 Introduction

The relief characteristics of drainage basins are of paramount geomorphological and hydrological importance; an importance reflected in a large number of descriptive relief terms (Drummond and Dennis, 1968). Drummond and Dennis (1968, p. 326), noted further that "inequalities in elevations of land surface occupy space and it is this situation which brings the question of area into the relief definition." The introduction of areal connotations into various relief measures has often been a source of confusion, much of it semantic. However, part of this difficulty is resolved when distinct areal systems such as drainage basins are studied. Therefore, determination of drainage basin relief parameters requires only that the descriptive measures be clearly defined and replicable.

Theoretically, relief measures reflect the potential energy of drainage systems (Strahler, 1952a, 1964). Sediment discharge is also closely related to basin relief (Schumm, 1956). In addition, these measures have value in comparative morphometric analyses insofar as regional tendencies may be ascertained. Again, a perusal of the literature, and especially of the article by Drummond and Dennis (1968), leads to the conclusion that no one relief measure is accepted as a standard for drainage basin studies. Therefore, as in the discussion

of basin shape measurement, several of the more commonly used methods have been applied here.

2:7:2 Property Definitions

Maximum basin relief, defined as the vertical distance between the mouth of the basin and the highest elevation in the basin (Strahler, 1952b) may be easily obtained from available topographic maps if the contour interval of these is not inordinately large. When combined with some measure of basin length to form a ratio this measure takes on greater meaning because of the dimensionless nature of that ratio. Accordingly, three ratios have been commonly employed. One of these was defined by Schumm (1956) as the ratio between the maximum relief of the basin and the longest dimension (in the same units) of the basin parallel to the main channel. Maxwell (1961) modified the relief ratio in using basin diameter, as defined for his analysis of basin shape. Thus, he re-defined the relief ratio as the ratio of basin diameter relief to basin diameter length (in the same units).

Melton (1957) employed a somewhat different measure which he called relative relief. This he defined as the ratio of basin relief to basin perimeter; basin relief being designated as the difference in elevation between the basin mouth and the highest point on the divide. Melton (1957, p. 5) expressed this measure as:

$$R = \frac{100 H}{5280 P} \%$$

where H equals the maximum height (feet) of the basin, from basin mouth to a point on the divide, and P is the basin perimeter (miles).¹⁴

¹⁴Some of the symbols have been changed here to conform with those used in this study.

The problem of minor perimeter crenulations unduly affecting the ratio has been cited (Maxwell, 1961) as a major weakness of this method.

In many studies of relief hypsometric analysis has been a focus of attention. Reviews of a number of related methods have been presented by Clarke and Orrell (1958), Clarke (1966), and King (1966) but here we are concerned with only one type, the percentage hypsometric curve. Since this measure is dimensionless it may be applied to basins of varying size. It was developed initially by Langbein et. al., (1947) but a more detailed treatment was undertaken by Strahler (1952b). Following is a summary of the percentage hypsometric curve method of drainage basin analysis, as developed by the latter author.

Calculation of two ratios is the initial step in construction of the percentage hypsometric curve. The first ratio is that of relative height; or, the ratio of the height of selected contours above the basin mouth elevation to the maximum height of the basin. The second ratio is that of relative area; or the ratio of the area between the contours selected and the upper perimeter of the basin to the total basin area. Figure 29 illustrates the method of coordinate plotting commonly used for presentation of the percentage hypsometric curve. Strahler (1952b, p. 1119) noted that this curve "expresses simply the manner in which the volume lying beneath the ground surface is distributed from base to top." He described three parameters which may be used to characterize natural curves in relation to model hypsometric curves. Leaving aside the mathematical expression of the curve-fitting method we may paraphrase Strahler (1952b) by stating

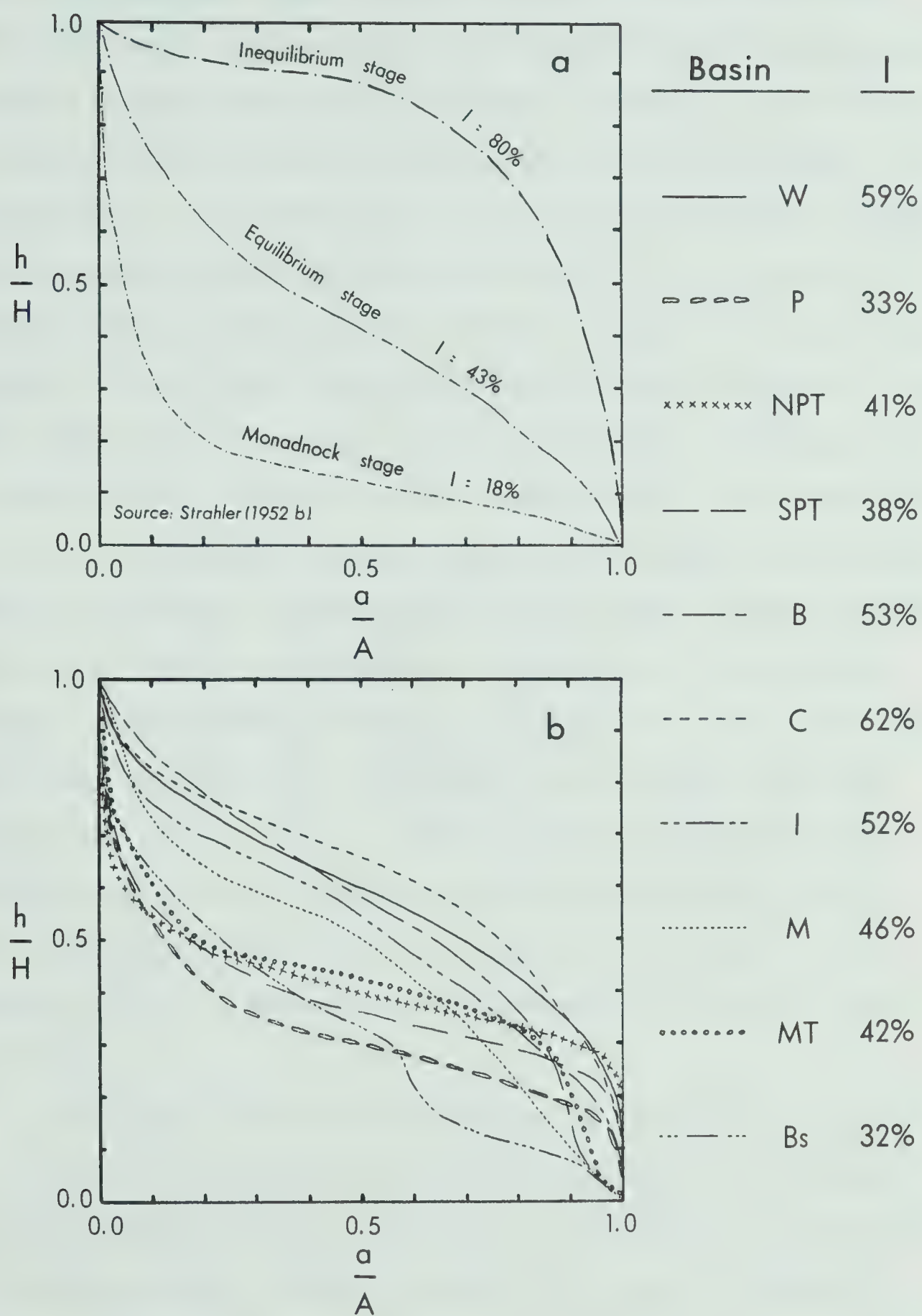


Figure 29 - (a) Model Hypsometric Curves
(b) Hypsometric Curves for the Study Basins

that the parameter r determines the hypsometric curve slope at the inflection point; the parameter z is an exponent indicating the general location of the curve; and the hypsometric integral (I) is the ratio of the area under the curve to the area of the entire reference square (Figure 29a). This integral was put forward as a measure of landmass volume remaining, on which erosion might act. It is possible to calculate each of these measures mathematically but the procedures are tedious. To facilitate ready comparisons of natural hypsometric curves with model curves most closely approximated Strahler (1952b) provided a number of model curves for graphic determination of the parameters r and z . The hypsometric integral may be quickly determined by planimetric or grid-square measurement of the area under the curve and calculation of this as a percentage of the total area of the reference square. In determining the value of r for a given natural curve the latter may be compared, by inspection, with the models constructed by Strahler (1952b, Plate I). The value of r is taken as that for the model curve family closest to the form of the natural curve. Then, knowing only the hypsometric integral and the value of r , the exponent value of z may be read off another graph (Strahler, 1952b, p. 1124).

While this fitting procedure cannot be considered as rigorous it allows rapid description of natural curves within the framework of the mathematically calculated models. On this basis it is possible to categorize natural curves into one of five families plotted by Strahler (1952b).

2:7:3 Central Alberta Data

Data calculated for the measures just outlined are contained in Table XI. These are, maximum basin relief (H) from basin mouth to some point on the divide, basin diameter relief (L_d), Schumm's (1956) relief ratio (R_1), Maxwell's (1961) relief ratio (R_2), Melton's (1957) relative relief (R_3), and the three hypsometric curve parameters, r, z, and I as defined by Strahler (1952b).

TABLE XI - BASIN RELIEF PROPERTIES

Basin	H (ft.)	L_d (ft.)	R_1	R_2	R_3 (%)	r	z	I(%)
Whitemud	535	535	0.0037	0.0037	0.137	0.25	0.38	59
Pipestone	615	535	0.0047	0.0041	0.160	0.01	0.28	33
N.P.T.	465	250	0.0057	0.0034	0.226	0.01	0.21	41
S.P.T.	515	480	0.0067	0.0062	0.227	0.01	0.22	38
Blackmud	290	260	0.0046	0.0042	0.141	0.25	0.41	53
Clearwater	320	305	0.0035	0.0034	0.126	0.25	0.30	62
Irvine	320	250	0.0035	0.0035	0.121	0.05	0.23	52
Maskwa	295	140	0.0042	0.0020	0.130	0.50	0.75	46
M.T.	395	340	0.0073	0.0063	0.183	0.01	0.20	42
Bigstone	495	430	0.0055	0.0048	0.195	0.05	0.42	32

It is obvious from Table XI that the central Alberta fourth order basins have small vertical dimensions considering their areal extent. This is also reflected in the dimensionless relief ratios, R_1 , R_2 , and R_3 . Schumm (1956, p. 612) cited several R_1 values for basins in diverse environments. Of seven areas studied only basins from the Gulf Coastal Plain and Hughesville, Maryland, areas had relief ratios comparable to the Alberta examples; the others exhibited markedly higher values. Schumm (1956) compared relief ratios with drainage density, stream

gradient, valley-side slope angles and basin shape (elongation ratio) and found some interesting relationships. With the exception of basin shape these measures were related to the relief ratio as positive power functions. Basin shape showed a weakly negative power relationship. These observations were tested for the local examples by plotting drainage densities and elongation ratios against the relief ratios on double logarithmic paper. The resulting coordinates appeared to be widely scattered, indicating no observable relationship.

Perhaps the most practical application of R_1 is in estimating sediment loss from drainage basins. It has been shown that the relief ratio and sediment loss are related in a semi-logarithmic manner (Schumm, 1956). Therefore, Schumm (1956, p. 613) stated that "once the characteristic regression trend has been established for a region the investigator may select areas of high potential sediment production from topographic maps." Sediment load measurements have not yet been initiated (to the writer's knowledge) for any of the local streams so it is impossible to state with any certainty that the ratio is useful for this purpose here. Two factors may, however, seriously inhibit such an application, assuming future sediment load measurements. First, because of the extremely variable temporal nature of runoff in local basins, very long-term data collection would be a prerequisite to establishment of a regional regression trend. Second, erosion processes and sediment discharge proceed at a very slow rate here and are undoubtedly related more to present land-use activities than to any single geomorphic parameter, such as the relief ratio. This notwithstanding, it would be most interesting to learn whether or not the relationship is

valid to any degree for this environment if future detailed sediment load studies are made.

The calculated values of R_2 and R_3 have been included mainly so that the comparative form of the Whitemud basin relief might be noted for various measures. Values of Maxwell's (1961) modified relief ratio differ markedly from the R_1 figures only for the North Pipestone Tributary and Maskwa basins and no further comment is necessary since the measure is basically the same as Schumm's relief ratio. Relative relief values were calculated by Melton (1957) for nearly 170 basins in Arizona, Colorado, New Mexico, and Utah and had a range between approximately 1 and 20 per cent.— The basins selected by Melton (1957) were all "maturely" developed and it was predictable, therefore, that the central Alberta values would be generally much lower than the American Southwest examples.

Figure 30 depicts the frequency distributions of the non-hypsometric relief values. Both absolute relief measures produced somewhat bi-modal forms, but in each case the Whitemud value falls in one of the modal classes. The R_1 values have a right-skewed distribution but again the Whitemud value may be considered reasonably typical since it occurs in the modal area of the histogram. Similarly, the basin emerges as typical with regard to R_2 and R_3 values. The latter distribution has the best-expressed modal class of any relief value distribution. Variation in the form of these frequency distributions may, then, be attributed mainly to the nuances introduced by the different methods of relief description.

It is opportune now to look more closely at the percentage hypso-

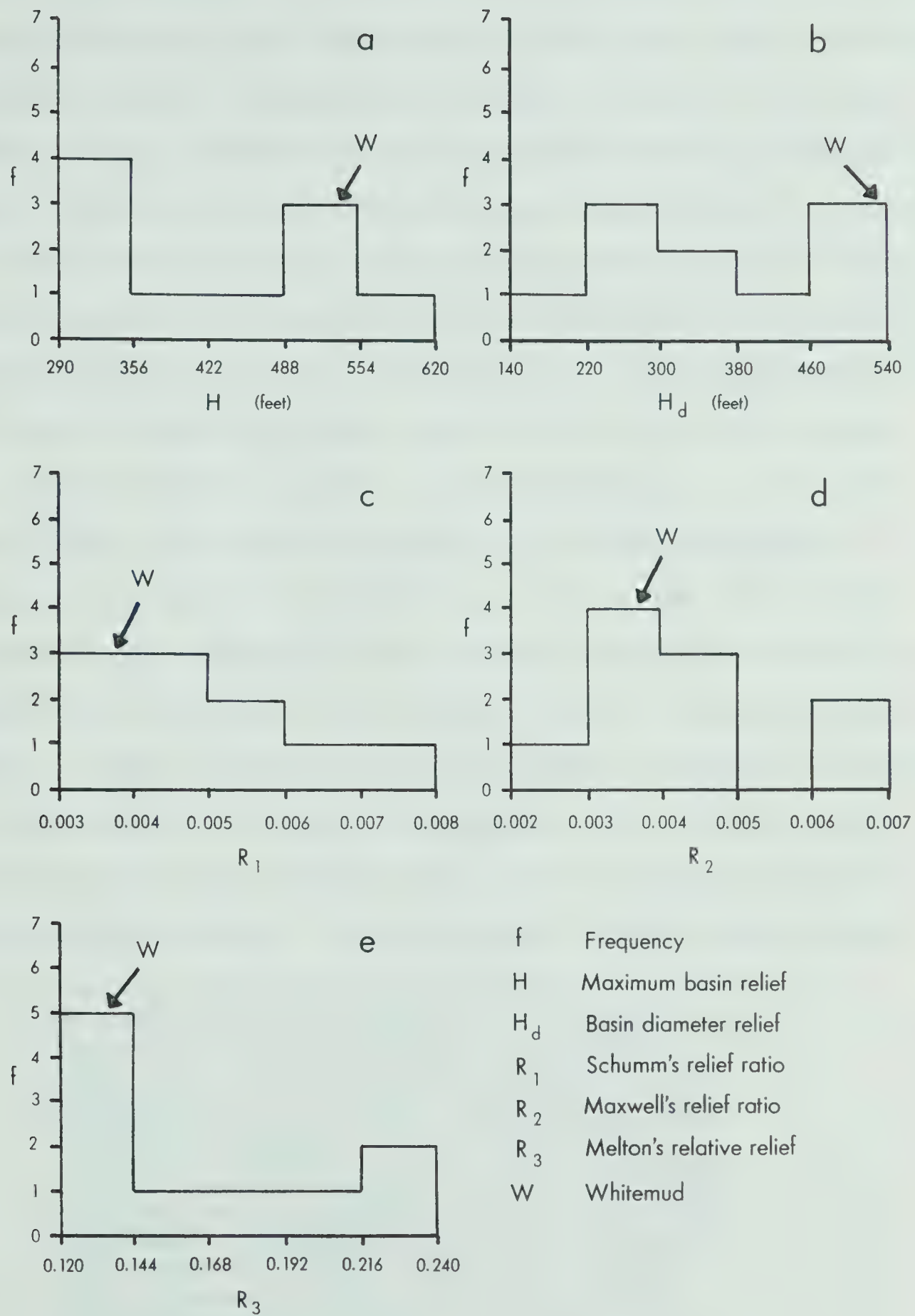


Figure 30 - Non-hypsometric Relief Properties

metric curve analysis. Great hopes were held by Strahler (1952), and since perpetuated in the literature (see: King, 1966), that this method might prove useful in evaluating the geomorphic history of drainage basins. To this end Strahler (1952b) analyzed percentage hypsometric curves for many basins at different stages of development (in terms of Davisian concepts of "youth," "maturity," and "old age."). He showed that relatively high (upwards of 60 per cent) hypsometric integrals obtain during the early stage of development. He designated this as the inequilibrium stage, where large portions of the initial surface are as yet unaffected by valley-side slope development. Coincident with "mature" (equilibrium) development the percentage hypsometric curves take on forms such that they pass close to the centre of the reference square (Figure 24a) with resulting hypsometric integrals between 35 and 60 per cent. An interesting feature is that (Strahler, 1952b, p. 1129) "in late mature and old stages of topography, despite the attainment of low relief, the hypsometric curve shows no significant variations from the mature form, and a low integral results only where monadnocks remain." Strahler (1952b, p. 1130), stated further:

From the standpoint of hypsometric analysis, the development of the drainage basin in a normal fluvial cycle seems to consist of two major stages only, (1) an inequilibrium stage of early development, in which slope transformations are taking place rapidly as the drainage system is expanded and ramified; (2) an equilibrium stage in which a stable hypsometric curve is developed and maintained in a steady state as relief slowly diminishes. The monadnock phase with abnormally low hypsometric integral, when it does occur, can be regarded as transitory because removal of the monadnock will result in restoration of the curve to the equilibrium form.

It was recognized, however, that this ideal sequence of events might be interrupted in various ways, with consequent anomalies in percentage hypsometric curves. Strahler (1952b) noted that in some instances regional up-warping and/or the rejuvenating effect of falling sea level might result in relatively high hypsometric integrals. Further, he showed that structural control may exert a marked influence on the position and form of the curves. For example, a plateau cut with deep canyons will produce a curve resembling in most respects the characteristic inequilibrium form.

With these reservations in mind we may now consider the local percentage hypsometric curves and integrals (Figure 29b and Table XI). Assume for the moment that these integrals reflect degrees of basin development, and further, that Strahler's (1952b) tentatively postulated transition points between inequilibrium and equilibrium forms (I equal to 60 per cent) and between the equilibrium and monadnock phases (I equal to 35 per cent) are valid. On this basis we could say that the Whitemud and Clearwater basins are close to the transition between inequilibrium and equilibrium, the North Pipestone Tributary, South Pipestone Tributary, Blackmud, Irvine, Maskwa, and Maskwa Tributary catchments are all in the equilibrium stage. It would be noted, further, that the Pipestone and Bigstone basins have just entered the monadnock phase. It is clear, though, that these integrals do not reflect varying degrees of basin development because the drainage networks are demonstrably "youthful." None of the basins can be considered "mature" because in general valley-side slope retreat has only just begun to encroach on the initial depositional surfaces. Equally obvious is the

fact that monadnocks (erosional residuals) cannot possibly exist in this area of Pleistocene sediments.

Therefore, we must consider the possibilities of regional up-warping, rejuvenation, and structural control in evaluating the local hypsometric curves. Bayrock (1964) has shown that the borders of Glacial Lake Edmonton have been tilted upward, towards the west, in the order of ten feet per mile during the last 11,000 years. He attributed this to isostatic readjustments of the Canadian Rocky Mountains because the magnitude and direction of uplift are at variance with the concept of isostatic response following the removal of large ice-sheets.

It is very doubtful that Holocene sea level changes have had any significant effect on these small, interior continental basins, so far removed from Hudson Bay. Good evidence of rejuvenation may, however, be found especially in the lower portions of basins tributary to the North Saskatchewan River. In postponing a detailed discussion of this evidence it suffices to note here that periodic climatic change seems the most probable major cause of local rejuvenation, aided perhaps by a certain amount of regional up-warping.

An element of structural control may also be discerned. This is particularly true of the Pipestone, North Pipestone Tributary, South Pipestone Tributary, Maskwa Tributary and Bigstone basins. The headward zones of these catchments are marked by relatively well-pronounced bedrock "highs." Farvolden (1963a), in describing the bedrock configuration of the area, named this bedrock region the Joffre Hills, composed of Paskapoo Formation sandstone and capped by Pleistocene sediments (Figure 28). He noted, also, that this phenomenon has the

surface expression of an escarpment. This structural aberration has been effective in producing hypsometric curves similar in shape to that observed by Strahler (1952b, p. 1139) for a basin containing a resistant butte rising above a shale lowland.

Therefore, the apparent existence of two distinct curve families (Figure 29b) may be explained quite simply as follows. Those basins exhibiting relatively high hypsometric integrals do not have abnormal elevations near their headward limits and, conversely, a structural escarpment or ridge exists at the heads of basins in the relatively low integral family. The anomalous bulge in the Bigstone basin curve (Figure 29b) is related to a subsidiary Joffre Hills escarpment located across the basin.¹⁵

Finally, one major criticism may be directed toward the percentage hypsometric curve method of analysis. In reviewing this type of approach Clarke (1966) observed that the area above the percentage hypsometric curve is equal to the volume removed by erosion and might be termed the erosion integral. For either the hypsometric or erosion integral to be meaningful in percentage terms of volume lost by erosion, or remaining to be eroded, we must make the often unwarranted assumption that the initial surface of the basin was nearly

¹⁵The reader may check the local values of r and z (Table XI) against Strahler's (1952) model curve graphs but in view of the fact that the local curves are determined in the main by structural and rejuvenative elements no further discussion of these parameters seems necessary.

or fully flat. A consideration of Figure 29a will indicate that a hypsometric integral of 100 per cent defines a perfectly flat surface and, therefore, a hypsometric integral of, for example, 60 per cent (calculated as a proportion of the reference square area) cannot denote the proportion of volume remaining if a sloping initial surface existed. If it were possible by some means to reconstruct the form of the initial basin surface, possibly through construction of contours from well-correlated erosion-surface remnants (Sharma, 1969) then a true approximation of volumes eroded and remaining could possibly be obtained. However, in most instances the reconstruction of initial surface configuration is not possible.

2:7:4 Summary

The preceding discussion of a number of relief measures has shown that the Whitemud basin may be thought of as typical of the local fourth order basins. Of note is the emergence of two families for the absolute measures (H and H_d) and also for the hypsometric curves. It is generally true that the existence of these two families may be explained in large measure by the presence of relatively resistant Paskapoo Formation sandstone toward the headward limit of some basins. These basins have relatively high absolute relief values and comparatively low hypsometric integrals.

Further, it has been shown that in this instance a number of disruptive elements such as structural control and rejuvenation have been operative throughout the geologically short-term development of the drainage systems. It is not possible, therefore, to use percentage hypsometric curve analysis to distinguish differences in

degree of erosional development in this locality.

2:8 Drainage Pattern

Descriptions of particular types of channel patterns (in the sense of plan geometry) have long been included in standard geomorphological texts (see: Strahler, 1951; Thornbury, 1954; Longwell and Flint, 1955). Perhaps the most comprehensive treatment of the descriptive terminology was that produced by Thornbury (1954) who noted that commonly encountered drainage patterns are termed dendritic, trellis, barbed, rectangular, complex and deranged. Dapples (1959) presented a somewhat different scheme, namely, longitudinal, advanced longitudinal, modified longitudinal, trellis, radial, dendritic steep slope and dendritic low slope. That such variations in terminology exist in the literature is an indication that the description of drainage patterns is almost always qualitative and subjective.

Problems of definition and recognition, therefore, usually forestall detailed analyses of pattern and a comprehensive treatment was not attempted in this study. However, in general terms, most of the central Alberta basins have channel patterns comparable to Thornbury's (1954, p. 125) deranged category. Describing this type Thornbury (1954, pp. 123-124) stated:

...there is usually [a] complete lack of structural and bedrock control. The preglacial drainage has been effaced, and the new drainage has not had time to develop any significant degree of integration. [This pattern] is marked by irregular stream courses which flow into and out of lakes and have only a few short tributaries. Much of the interstream area is swampy and frequently the streams are mere threads of water through swampy areas.

A major exception, however, is that of the Whitemud basin channel

network. This closely resembles the advanced longitudinal pattern suggested by Dapples (1959, p. 190), at least in the arrangement of the major third and fourth order channels (Figure 6). Again, it would appear that the best explanation of this phenomenon lies in a consideration of the preglacial bedrock configuration. A well-defined preglacial valley existed almost exactly parallel and coincident with portions of the present main Whitemud valley. It may be noted (Figure 2) that a large proportion of the Whitemud basin area consists of residual material (where bedrock is at or close to the surface) and lacustrine sediments mantle much of the lower half of the basin. As a result, prominent ridges of moraine, or other types of deposit, occur only occasionally here. The bedrock configuration has, therefore, probably played a major role in determining the gross shape and slope components of the present basin. Thus, a relatively well-integrated postglacial drainage system has evolved because glacial erosion and deposition failed to obliterate the former well-developed valley.

Although the drainage patterns of the other basins reflect some degree of adjustment to preglacial bedrock influences, and especially to Pleistocene spillway channels, the Whitemud catchment is singled out as having the greatest degree of pattern integration.

2:9 Summary of Fourth Order Basin Morphometry and Conclusions

Two primary objectives have been the aim of this chapter. The first was to present data which might adequately describe, in non-qualitative terms, the form of some central Alberta fourth order drainage basins. Allied to this, the second aim was to evaluate the Whitemud basin for typicality in the context of local morpho-

metric values. It was pointed out that statistically sophisticated comparative techniques could not be applied, largely because of the small number of basins studied, but also because of the non-random method of basin selection. However, it was felt that some measure of deviation from the modal class of each frequency distribution might best describe the degree of Whitemud typicality in each case. To this end a simple ranking procedure was set up. The modal class (assumed to be the best measure of central tendency) was designated by the figure 0 and positive or negative deviations away from the modal class were enumerated as +1, +2, ... -1, -2 ...respectively. Obviously, the resulting array of values (Appendix A) describing degree and direction of deviation from the modal class can only be interpreted as a very general descriptive device. It cannot, for example, be easily adapted to show the relative strength of the modal class of any given frequency distribution. In other words, the modal class may be defined as such on the basis of having merely one more frequency of occurrence than any of the other classes, even if a comparatively regular distribution of the frequencies obtains throughout the range. This categorization is weakened further when bi-modality exists. Accepting the fact that such weaknesses do present themselves we may, nevertheless, use this as a simple comparative tool. Perhaps the greatest value of this ranking procedure is that a mass of morphometric data may be integrated into an easily comprehensible tabular, or preferably, graphical form. In this study the tabulated values appear in Appendix A, and for convenience a graph (Figure 31, along with Table

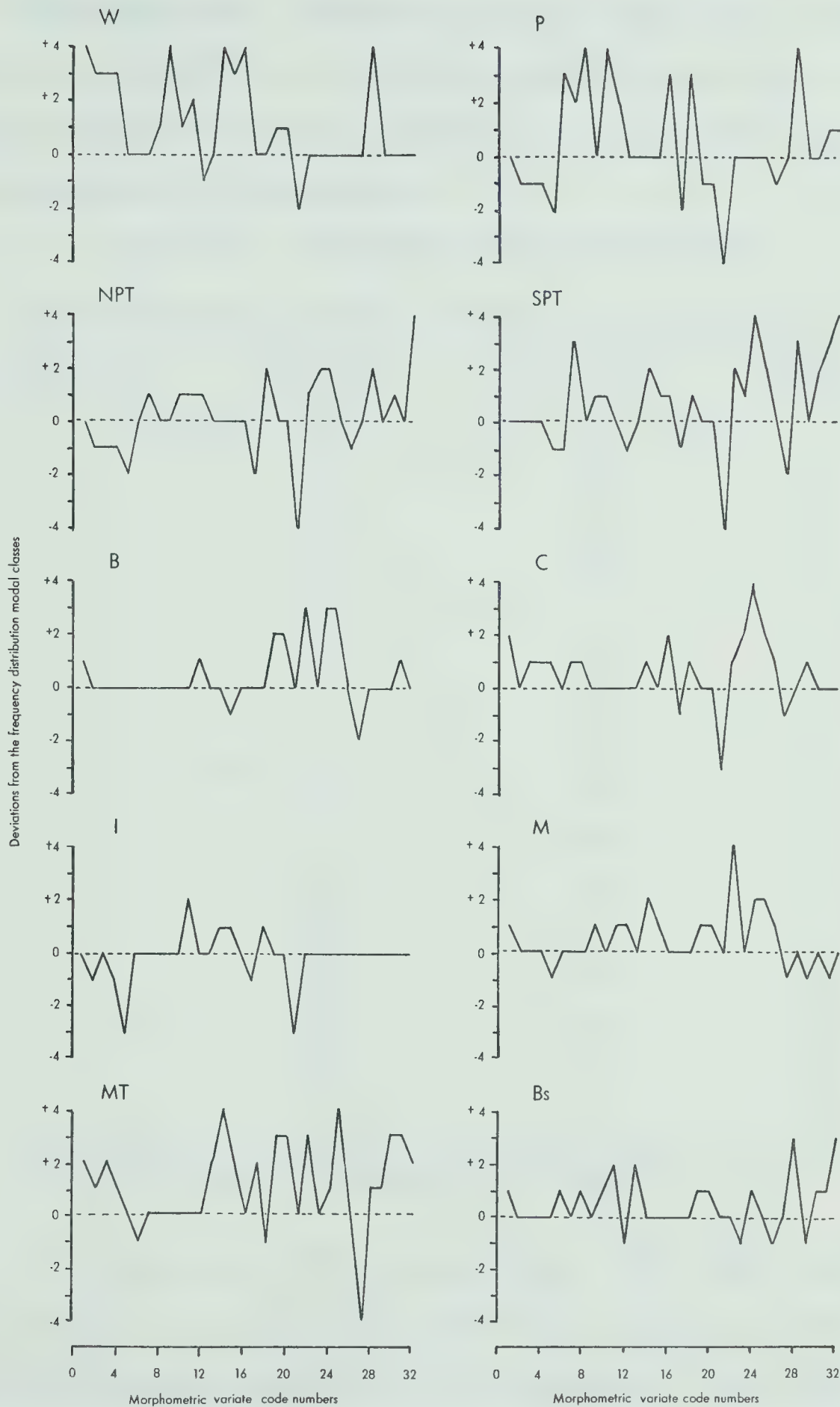


Figure 31 - Diagrammatic Comparison of the Morphometry of the Study Basins

XII), is included here. It may be noted that values deviating from the modal class often fall on the positive side of the modal rank. This is because many of the frequency distributions of morphometric properties have some degree of right-skewness.

TABLE XII - CODE NUMBERS/MORPHOMETRIC VARIATES

Code Number	Variate	Code Number	Variate
1	N_1	17	D
2	N_2	18	C
3	N_3	19	F
4	N_k	20	F_1
5	%E	21	D_1
6	$\bar{L}_1$	22	S_1
7	$\bar{L}_2$	23	S_2
8	$\bar{L}_3$	24	S_3
9	$\bar{L}_4$	25	S_4
10	$\bar{L}_k$	26	k
11	$\bar{G}_1$	27	K
12	$\bar{G}_2$	28	H
13	$\bar{G}_3$	29	H_d
14	$\bar{G}_4$	30	R_1
15	$\bar{G}_k$	31	R_2
16	A	32	R_3

(To be used in conjunction with Figure 31)

If we consider first, for each basin, the total number of times values fall in modal classes three broad groups of basins, synonymous with decreasing typicality, may be demarcated. At one end of the spectrum the Irvine and Blackmud basins appear the most typical, with

modal rank scores of twenty-three and twenty respectively, out of a possible thirty-two. Well below these scores are those of the second group, consisting of the Whitemud, Maskwa, Bigstone, Clearwater, North Pipestone Tributary and Pipestone catchments. Scores for these basins range from sixteen (Whitemud and Maskwa) down to thirteen (Pipestone). The least typical group is made up of the South Pipestone and Maskwa tributary basins, both with a score of eleven. On this basis we could assert, with some confidence, that the Whitemud catchment is aberrant in relatively few aspects. However, this conclusion should be somewhat tempered on inspection of scores summed for the modal and two immediately adjoining positive and negative classes. Such a summation does not yield three distinct groups and, furthermore, places the Whitemud basin lower in the typicality order. In this case scores range from twenty-nine to eighteen. Six basins have higher values than the Whitemud basin which tends to have markedly atypical absolute and dimensional values, such as the number of channels in a given order. Of the ten basins, only the Pipestone catchment closely parallels the Whitemud in this regard. Since dimensional values ordinarily exhibit wide ranges of values for basins of a given order we are reassured that in most respects the Whitemud basin is reasonably representative of the local study area.

The interrelationships of geology, soils and morphometry are by no means clear-cut in this environment. Certain obvious correlations do present themselves, for example, the general absence of channels in areas of aeolian sand deposits. The Pipestone basin is a prime example of this. Apart from such direct relationships

there seems to be little consistency among other factors. In some areas till is marked by relatively high drainage densities, in others the opposite is true. The Maskwa Tributary and Pipestone basins illustrate this. Soil/morphometry relationships are also obscure and it would appear that even the solonetzic zones are not particularly influential.

It bears repeating that the preglacial bedrock topography has played a part in determining certain morphometric characteristics of the basins. However, the vagaries of glacial and glacio-lacustrine deposition, and certainly the formation of a number of spillways during the existence of Glacial Lake Edmonton, have also been operative. As a result it is almost impossible to assign degrees of importance to particular physical factors for any given basin. The emergence of the Blackmud and Irvine basins as the most typical, in crude terms of central tendency, is illustrative of this point. These basins appear to have in common only the present distribution of soils, and possibly, the preglacial bedrock configuration. Basic differences in surficial geology, and the existence of the Gwynne Channel in one, confuse the issue. This notwithstanding, the basins are remarkably alike, except in terms of shape.

Before concluding this description of fourth order basin morphometry we should briefly review the present status of morphometric theory. Allusions have already been made to a dichotomy of opinion as regards the so-called "Horton analysis" of channel networks and associated basin form properties. The apparently divergent opinions present in the recent literature are based on two views of natural evolutionary processes. On the one hand are investigators

(see: Woldenberg, 1966; Strahler, 1969) who consider that the regularity and general applicability of the laws of network and basin form reflect orderly development through time. This view was adopted earlier by Horton (1932, 1945). Recently, however, a number of writers (Scheidegger, 1967, 1968b, 1968c, 1968d; Smart, 1967, 1968; Smart and Surkan, 1967) have extended the arguments introduced by Leopold and Langbein (1962), Milton (1966), and Shreve (1966) that these developmental laws are in fact statistical phenomena and that basin development is explained better in terms of probability theory of stochastic processes. For example, Smart (1967) used probability theory to suggest that stream number/order plots for low order basins will have a greater tendency to be concave upward and that with increasing order the approximation to a straight-line, semi-logarithmic fit will improve. However, Smart (1967, p. 776) concluded that:

just what [these] results imply concerning the validity of Horton's law [of stream numbers] is to a certain extent a semantic question. Neither Horton nor later writers on the subject, with the exception of Shreve (1966), have been very explicit about the nature of the deviations that might be expected in real systems. It seems clear from the context of the papers, however, that many people have implicitly interpreted the law as a literal statement about the central tendency of stream numbers in a hierarchy of stream systems. Our results show that any attempt to apply the law in such a strong sense leads to internal inconsistencies. Horton's statement can also be interpreted in a much weaker sense, namely, simply as an approximation that becomes successively better as the order of the basin increases.

In a paper summarizing the two divergent views of network formation Scheidegger (1968c) pointed out that the growth-model concept (specifically, the allometric growth idea presented by Woldenberg (1966)

and supported by Strahler (1969)) can apply, theoretically, to only one type of network, that which is "structurally Hortonian." By this is meant networks having identical bifurcation ratios for all sub-basins of lower order within, and for, a main basin.¹⁶ However, many natural basin networks do not satisfy this condition, but at the same time conform to Horton's law of stream numbers. The graph-theoretical approach advocated by Scheidegger (1968c) apparently overcomes this difficulty in accounting for both types of network. The mechanics of the related statistical arguments indicate that random development of networks is more probable than systematic growth.

In sum, then, we are caught in something of a theoretical cleft-stick. The theorists themselves seem loath to reject Horton analytic methodology, notwithstanding the fact that basic differences of opinion exist. Until these arguments are resolved, then, the non-theorist is justified in using and looking upon these methods of morphometric analysis as a relatively efficient means of making erosional landform inventories and classifications. From these he may at least reach certain conclusions regarding regional geomorphology, if not clarifying the theoretical formulations. That such an approach is valuable has been demonstrated in the present study, wherein the White-mud basin and channel network form was ranked in a hierarchy for the immediate region.

¹⁶Scheidegger (1968c) specifically used the Strahler channel ordering system in developing this argument. Note also that "all sub-basins of lower order" does not include first order basins which, by definition, do not have bifurcation ratios.

CHAPTER THREE

INTERFLUVE, VALLEY AND CHANNEL CHARACTERISTICS OF THE WHITEMUD BASIN

3:1 General Introduction

The following detailed description focuses on aspects of the Whitemud basin geomorphology most intimately related to past and present hydrological conditions. The preceding morphometric analysis served as a quasi-regional base whereby the channel network morphology and basin parameters of relief and shape were established.

The interfluve areas were described in terms of maximum slope angles, slope lengths, and slope orientations. Valley and channel landforms, including terraces, valley-wall slopes, riffles and channel cross-sections, reflect the most active geomorphic elements of the past and present fluvial systems. In particular, valley and channel landforms offer the best opportunities for field observation, something which is not so readily facilitated by methods of morphometric analysis. As Leopold, Wolman and Miller (1964, p. 479) stated:

Geomorphology is, or should be, primarily a field study, and the field aspects cannot be eliminated by any growth in availability of maps, aerial photographs, or computers. These still are tools to assist in the pursuit of field investigations.

3:2 Interfluvial Characteristics

3:2:1 Introduction

The local area has a very coarse-textured erosional terrain and the depositional surfaces produced during and after the last Pleistocene glaciation are largely unaltered (Bowser, et. al., 1962). Valley-side slopes are clearly in an early stage of evolution. Little research into current regional slope processes has been undertaken, except for an erosion-plot study initiated in 1949 by the Soils Department, the University of Alberta. This site was established on virgin grassland at St. Albert and the project was directed toward the correlation of rainfall intensity, runoff, and soil loss. Two progress reports (Toogood and Newton, 1955; Toogood, 1963) indicate that recent sheet and rill erosion have been minimal. Gullying may occur under certain conditions but this is more a function of human interference through the disruption of vegetation than of natural environmental conditions. In the main, the interfluvial slopes have probably remained relatively constant in form for much of Holocene time.

It might be expected that the Whitemud basin interfluvial surfaces would vary directly in form with the type of surficial geology. As a tentative formulation it was considered that areas mantled with glaciolacustrine and fluvial sediments (Bayrock and Hughes, 1962) should have relatively low slope angles. Toward the basin margins, and especially near the headward limits of the catchment, steeper slopes should prevail. Two factors make this a reasonable supposition; first, the higher ground is more liable to be covered with moraine than lacustrine

material, and second, the basin margins are probably closely coincident with upstanding portions of the preglacial terrain.

With respect to interfluvial slope orientations two general formulations were considered. First, if the assumption that erosion processes of slope development have been generally ineffective is valid then a preferred orientation (if any) must be attributed to some other factor. The existence of a preferred slope orientation (or orientations) is likely to be a product of both the preglacial form and distribution of slopes and, possibly, in part, of glacial or glacio-lacustrine depositional processes. Second, the gross pattern of ground slope in the basin (as indicated by contours in Figure 6, and especially on the pertinent National Topographic Series, 1:50,000 maps) exhibits an approach generally tangential to the major channels. However, this may be misleading in indicating that overland flow theoretically could proceed from near the basin divide to the main channels. A sample of slope orientations taken at true scale in the field should reveal whether this gross tendency is reflected on the ground.

3:2:2 Methodology

To accumulate a slope sample a grid-square method of fixing point locations was adopted. Points were selected by randomly placing a grid, divided into one square mile units, over a map of the basin. The grid-square intersections were taken as the sample points which were then fixed on the map to be used in the field. Sites ostensibly located on roads, buildings, railroads and other grossly unnatural surfaces were ignored. Of the 120 points selected only two had to be abandoned; one was too close to a railway embankment, and, access to

the other was denied. Three measurements were taken at each site:

(1) the maximum slope angle; measured with an Abney level, to the nearest one quarter of a degree, between two staffs, 5 feet long, placed at each end of the demarcated slope

(2) the slope azimuth; measured with a Brunton compass to the nearest 10 degrees, correct to true north

and (3) the slope length; measured by pacing between discernible breaks of slope.

The Abney level is inefficient at very low slope angles. Some correction of the readings was obtained by sighting both up and down the selected slope and averaging the values if they differed. Because very low slope angles are characteristic here readings were taken to the nearest one quarter of a degree on the Abney level. While this necessarily promoted extreme care in taking each measurement the inherent qualities of the instrument make it doubtful that such a degree of accuracy can in fact be obtained. It was thought better to err this way, however, than to take rough measurements. The maximum slope angle at most locations was determined from a number of alternative measurements, eliminating a certain degree of subjectivity. A further problem was that often the same slope angle extended over a broad sector from an obvious high point, or apex. The average azimuth was accepted in such cases.

Slope length measurement introduces the greatest element of subjectivity as breaks of slope are often difficult to recognize in areas of gentle relief. However, errors are probably not more than

5 yards for the longer slopes, less for the shorter ones. Operator error, or personal bias, was minimized in that only the writer delimited the slope lengths, azimuths, and angles.

Strahler (1956) has shown that the grid-square method is satisfactory for estimating the population mean maximum slope angle from a sample. He pointed out, though, that a truly random sample must be employed if an estimate of the population variance is required. Since very low-angle slopes predominate here, and no accurate estimate of possible measurement errors could be obtained, it was thought unnecessary to estimate the variance.

3:2:3 Whitemud Basin Data

A general impression of gross slopes in the basin may be gained from an inspection of Figure 32 which depicts east-west profiles of the catchment, separated by intervals of two miles. It is apparent that these profiles do not represent true slopes since the diagram has a large vertical exaggeration and the profiles were not constructed orthogonal to the contours. Nevertheless, the relatively flat area marked by Glacial Lake Edmonton lacustrine sediments emerges clearly, as do a few smaller areas with relatively steep slopes. Figure 33 illustrates the well-defined depression, probably indicative of the preglacial Whitemud valley centre line, into which the present Whitemud Creek is incised. This locality is approximately coincident with that portion of the valley shown in cross-section J, Figure 32.

The distributions of maximum slope angles and slope azimuths measured in the Whitemud basin, as well as the basic surficial geology

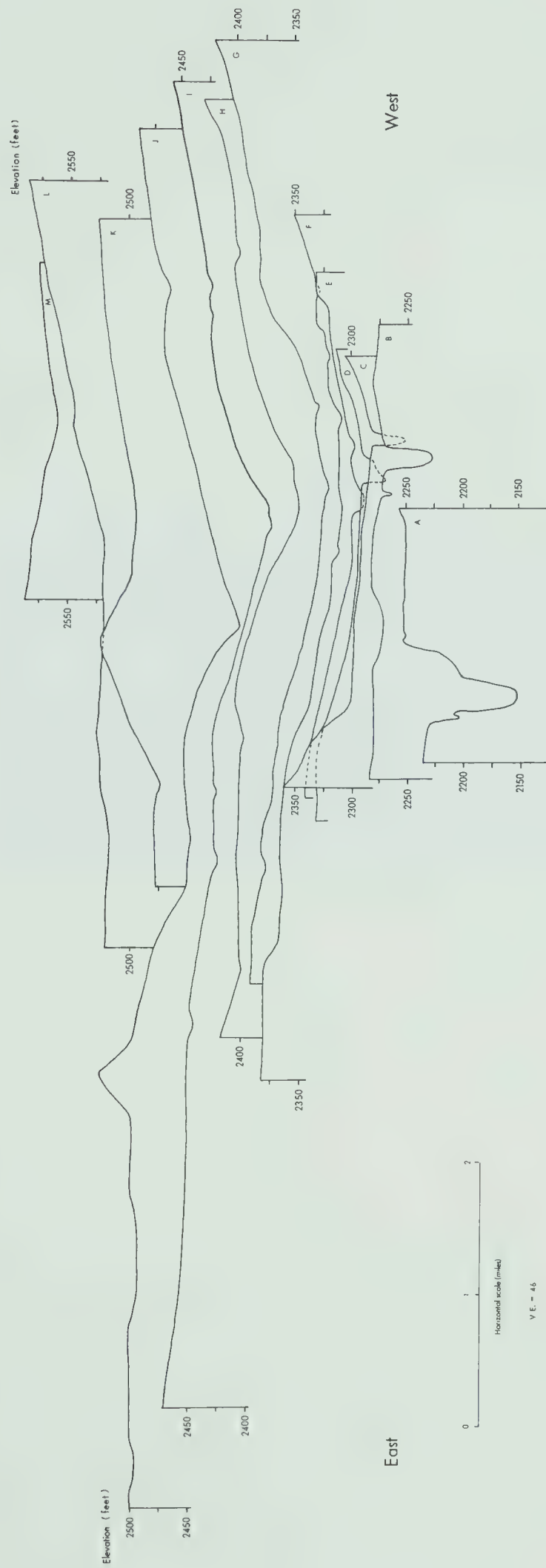


Figure 32 - Cross-profiles of the Whitemud Basin

a

Figure 33 - This depression, into which part of Whitemud Creek is incised, coincides with the thalweg of a preglacial valley.



units, are shown in Figure 34.¹ It was assumed that maximum slope angles are in all probability most intimately related to the nature of the surface materials. For the present purpose the surficial geology was simplified into two basic units — lacustrine, and non-lacustrine. This assured that approximately half of the sample slope angles fell in each zone. A more specific subdivision of the surficial geology would have resulted in widely varying sample sizes, to the detriment of mean maximum slope angle calculation.²

A slightly different approach was used for the analysis of slope orientation. Here the basin was divided into halves by extending the main, fourth order channel to the divide. This also assured that approximately half the sample points fell in each section. The rationale for selecting basin halves rather than different surficial geology zones is as follows. It was considered that the tangential nature of the gross interfluvial slopes is probably more a function of the preglacial valley imprint on glacial and glacio-lacustrine deposition than of intrinsic qualities of these sediments. As a check, however, the frequency distribution of slope orientations was also plotted for the two geologic units.

¹It should be noted that interfluvial slopes were not sampled in the area where the Edmonton International Airport is located, immediately north-west of Leduc (Figure 1).

²Arrows in Figure 34 are directed downslope from each sample point location. The length of these arrows could not be readily adjusted to represent slope length and have, therefore, been kept standard to facilitate cartographic presentation.

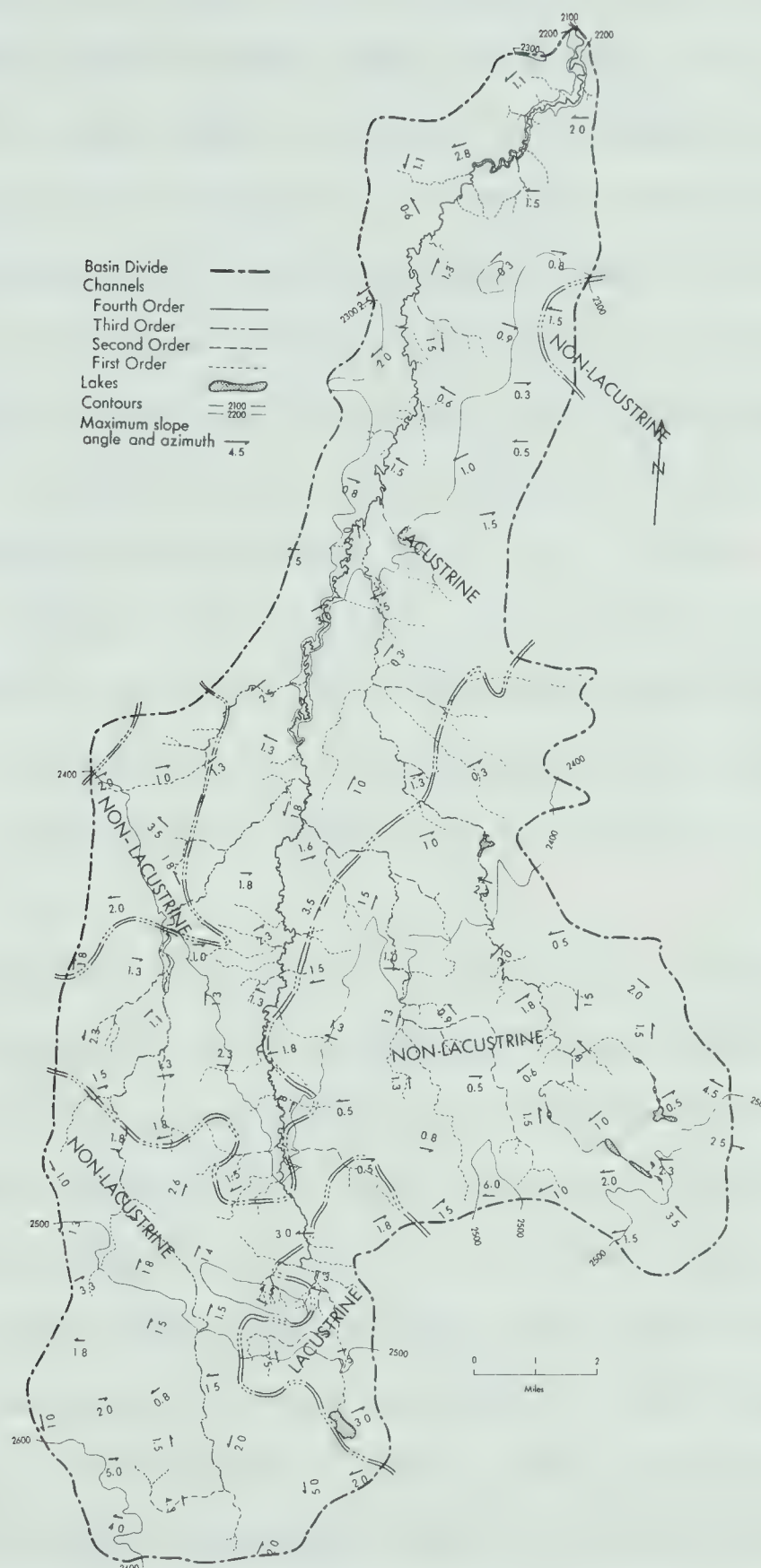


Figure 34 - Interfluvial Maximum Slope Angles and Azimuths in the Whitemud Basin

It is apparent from Figure 35a that the majority of interfluvial maximum slope values are lower than about 2.50 degrees. The right-skewed distribution stems essentially from a few uncommonly steep slopes which occur in the zones of till or residual surficial geology (Figure 2). As expected, the lacustrine sediments have relatively low-angle slopes in the basin context. However, the mean maximum slope angles do not vary greatly from lacustrine to non-lacustrine materials but the lacustrine values have a comparatively small range and dispersion around the mean (Figure 35b and c). In terms of Savignier's (1965) and Bunting's (1967) criteria the interfluvial slopes may be considered, mainly, as "flat." While the slope length data are rather suspect because of difficulties in field demarcation it is tentatively suggested that, on the average, lengths of 20 to 25 yards are typical (Figure 35d).

Methods of representing preferred orientations have been developed particularly for studies of till coarse fraction fabrics (Holmes, 1941). Rose diagrams are often used to graphically illustrate the till data and here these have been employed for slope orientations (Figure 36a and b). Although field measurements were taken to the nearest 10 degrees the data were further generalized into nine equal sectors of 40 degrees each. This was essentially an expedient measure to enable application of the Chi-square significance test (Andrews, 1963). One requirement for this test is that each class should have an expected frequency of at least five (Blalock, 1960). Grouping into 40-degree sectors satisfied this requirement. The distributions in both the west and east halves of the basin were shown by this test to have preferred orien-

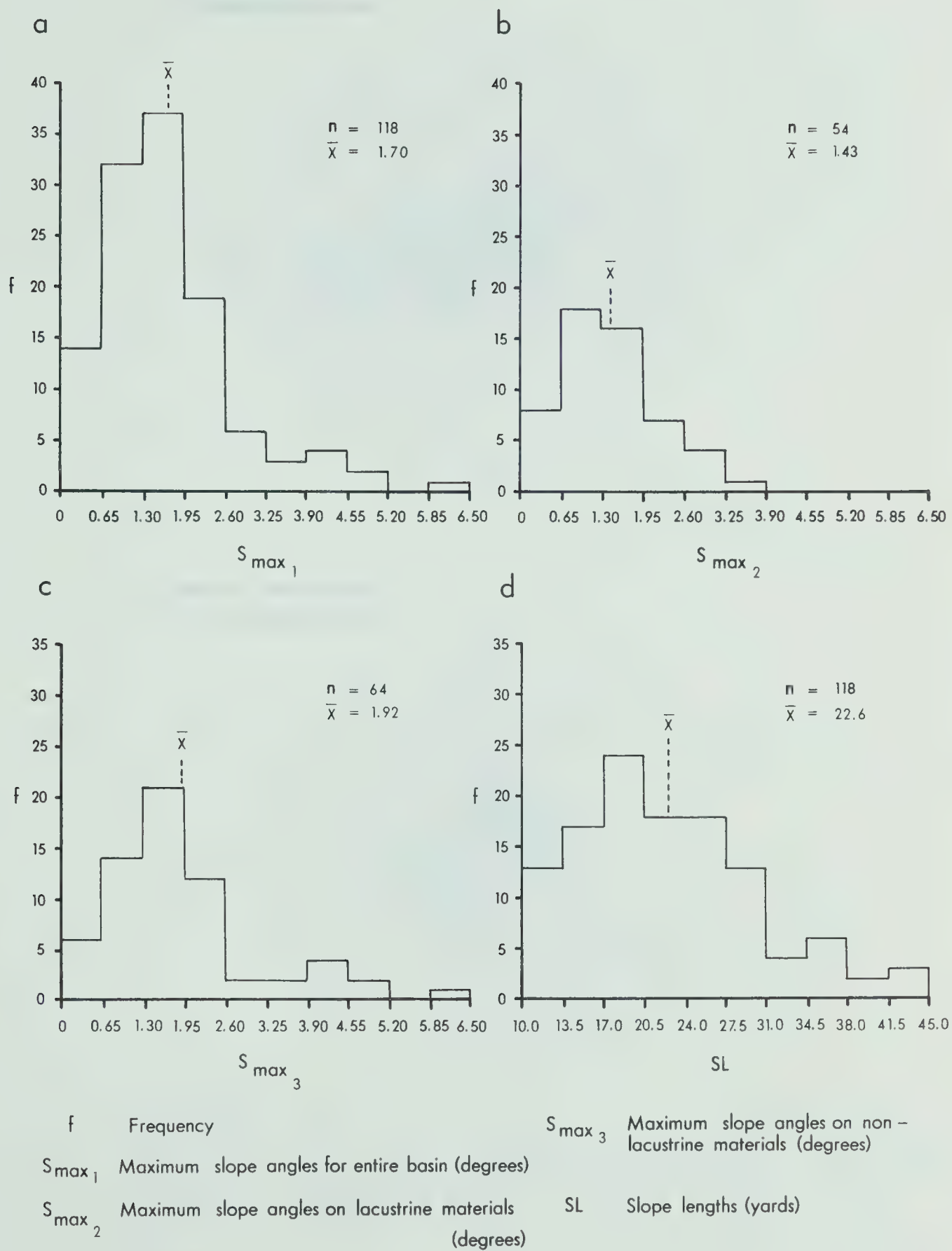
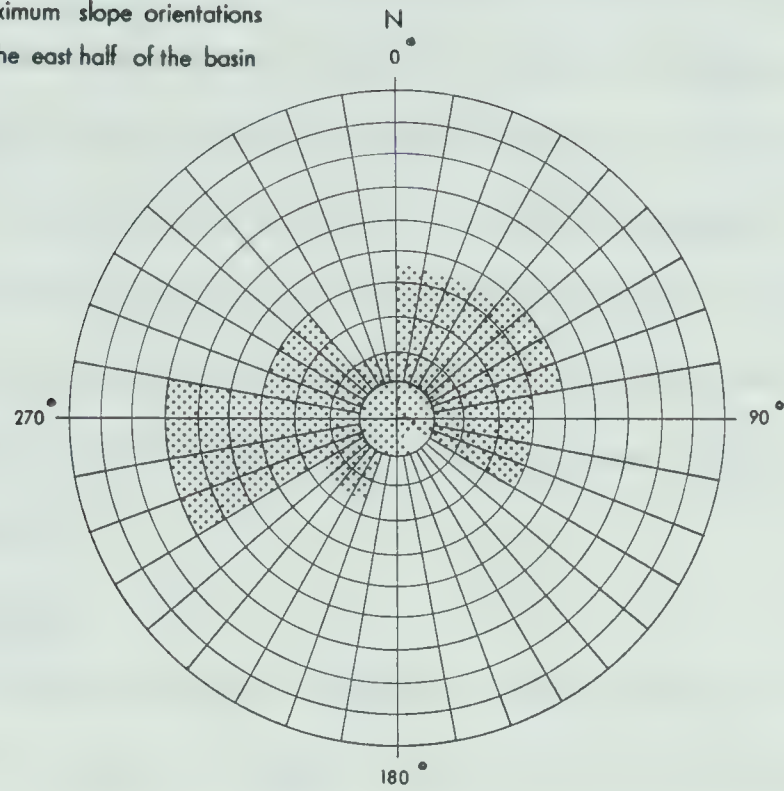
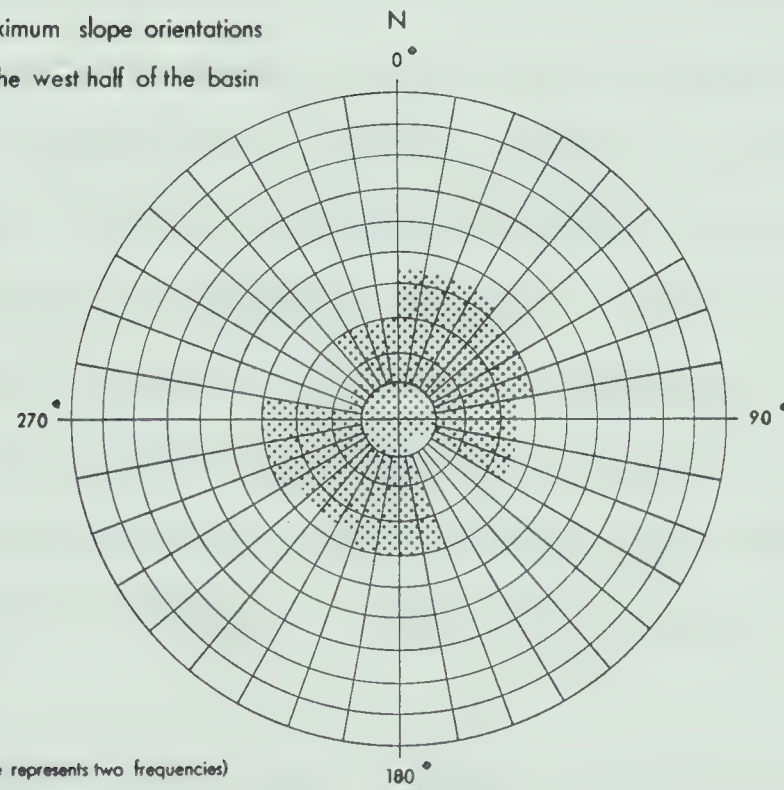


Figure 35 - Interfluvial Maximum Slope Angles and Slope Lengths

a Maximum slope orientations
in the east half of the basin



b Maximum slope orientations
in the west half of the basin



(Each circle represents two frequencies)

Figure 36 - Orientations of Interfluvial Maximum
Slope Angles

tations at the 5 per cent significance level. However, at the 1 per cent level the null hypothesis (that the observed frequencies do not differ significantly from the expected frequencies) was accepted for the west half. This notwithstanding it seems reasonable to assume that preferred orientations do exist for the maximum interfluvial slopes, if weakly in the western half of the basin. In both cases (Figure 36a and b) the north-west and south-east azimuths have a paucity of observed frequencies.³

3:2:4 Summary

Insofar as the Whitemud interfluvial slope forms are concerned the following generalizations are made. First, it would appear that the gross surface slopes, which tend to converge at a tangent to the main Whitemud valley, are not necessarily reproduced at specific sites on the interfluvies. In other words, extreme lengths of overland flow are not possible (even in the rare cases of sufficient moisture being available for runoff over such distances) because of minor surface undulations and depressions which counter the gross slope components. Maximum slope angles exhibit some variability between types of surficial geology but, apart from a few in the range of 5 to 7 degrees, may in general be considered as flat (< 2 degrees) or nearly so. If gross slopes are re-

³Although not included here in graphic form the lacustrine and non-lacustrine slope orientation frequency distributions showed that the lacustrine slopes have a weak preference for azimuths between 0 and 120 degrees. Non-lacustrine materials exhibited a strong preferred orientation between 240 and 280 degrees, with a secondary weak preference in the range of 0 to 80 degrees, closely paralleling the east-half distribution (Figure 36b).

flected by slopes at specific sites then the most probable preferred orientations should fall approximately in the north-west and north-east sectors. However, such is not the case since the north-west is distinguished by an absence of observed frequencies. It is suggested that the preferred orientations observed might in some way reflect the direction of Pleistocene ice-sheet advance. Both the north-east and north-north-west directions have been cited as possible source azimuths (Gravenor and Bayrock, 1955; Gravenor and Ellwood, 1957; Westgate, 1969). If steep depositional slopes formed normal to the direction of ice advances (for example, on the proximal and distal sides of ice-push ridges) these may have influenced the interfluvial surface micro-slopes. It has been noted elsewhere that till-ridges, flutings, and other protuberances were often poorly concealed by subsequent glacio-lacustrine deposition (Gravenor, Green, and Godfrey, 1960). This appears to be the most likely explanation for the preferred orientations observed but the topic requires further investigation.

3:3 Valley Characteristics

3:3:1 Introduction

The main Whitemud valley reveals most clearly the historical development of the basin drainage system. In the headward portions of the basin, channels are only faintly etched on the interfluvial surfaces, and valley and channel are essentially synonymous. Matched terrace surfaces exist in the greater length of the main valley, and in the lower reaches of major third order tributary valleys. The terrace surfaces were mapped in detail and most exposures of the alluvial deposits were described and measured. Certain sections (exposures) were sampled so that

the size distribution of the sediments might be determined and compared for the various terraces.

A secondary concern was the nature of valley-wall slopes in the main Whitemud valley. Prior to slope sampling it was postulated that:

- (1) valley-wall asymmetry should be minimal because the strike of the valley is almost north and south, thus reducing micro-climate differences
- (2) valley rejuvenation should be reflected by relatively steep slopes near the basin mouth

and (3) if processes of slope development have been at all active, evidence of slope angle transformation should be revealed for older slopes, that is, those slopes most removed in space and time from basal stream sapping.

Although sampling of both terrace alluvium and valley-wall slopes was limited by time considerations, and the large field area involved, the data provide basic comparisons and point to areas of interest for future more intensive research.

3:3:2 Methodology

Large-scale maps (1:4,800) with a contour interval of 4 feet, were prepared by photogrammetric methods⁴ for the Whitemud valley, from the Blackmud Creek junction to the Leduc/Calmar highway (Figure 1). This small contour interval, based on geodetically determined control ele-

⁴Mapping was done by the Technical Division, Department of Lands and Forests, Government of Alberta, and based on aerial photography, of scales 1:4800 and 1:12,000, flown by Range Aerial Ltd., Calgary, in May and June, 1967.

vations, allowed rapid and accurate checking of the elevations of the field-mapped terrace surfaces. It is usual practice to combine the use of altimeters and barographs for terrace height surveys (Debenham, 1955) but this is a very time-consuming process, and the value of such special-purpose mapping as employed in this study is manifest.

The larger-scale aerial photographs were invaluable for plotting of field observations on transparent overlays. Stereoscopic examination of these also allowed efficient use of time available in that promising sites could be located prior to the field survey.

In the field, terrace surface remnants were delimited and plotted on the aerial photograph overlays and each exposure of alluvium was described in the following manner:

- (1) The major vertical dimensions were measured above the bottom of the present Whitemud channel adjacent to each section and, where possible, the upper limit of bedrock was demarcated. Above this, recognizable units of the alluvial sequence were measured and briefly described.
- and (2) Sections were sampled for mechanical grain-size analysis. It was thought unnecessary to sample gravel and where present this was merely described in terms of general lithologic composition and place in the alluvial sequence. Therefore, two to four samples of fine-grained sediments were taken from the selected sections. Where very different strata were apparent one sample was taken from each. Alternatively, if the vertical distribution of material seemed relatively homogeneous two or three equally spaced samples were taken in the

vertical plane, depending on the depth of the particular bed.

The grain-size analysis was carried out using sieve and pipette techniques. Briefly, the pipette method allows measurement of the percentage clay weight for a given sample; sieving allows calculation of the sand percentage weight; and the silt component is then obtained by subtraction. Robinson (1922) outlined the early development of the pipette technique and Folk (1961) described one type of pipette analysis. The method employed here was developed primarily for use in soil studies⁵ and has the major advantage of being a rapid means of grain-size determination. A minor disadvantage is that because only the basic clay, silt, and sand percentages are obtained it is not possible to construct cumulative curves from which such statistics as the mean, standard deviation and median diameters may be read for individual samples. If such data are required it is necessary to use the hydrometer method or the modified pipette technique described by Folk (1961). Since the main concern was with simple comparisons of the various terrace deposits, the more rapid and general pipette method was adopted.

In order to evaluate the three valley-wall slope postulates a series of stratified random sample points were established. The main valley, as depicted on the large-scale maps, was divided into four equal sections from the basin mouth to the Leduc/Calmar highway. Thus, along

⁵Dr. R.B. Bryan, Department of Geography, University of Alberta, kindly instructed the writer in the use of these techniques, outlined in Appendix B.

each of the two valley walls four equal segments were created. Within each segment twenty-five points on the valley-wall rim were located in the following way:

- (1) A random numbers table (Krumbein and Graybill, 1965, p. 445) facilitated a series of map measurements (inches), successively, along the rim from the downstream end of each segment.
- (2) The figure 0 in the random numbers table was taken to equal 10 inches.
- (3) The sampling was without replacement so that any site could not occur more than once.
- (4) If in the first run of numbers, twenty-five points were not accounted for in any segment, the procedure was repeated until that many were included.

and(5) Areas disrupted by coal-mining, and zones where major tributaries enter the valley, were ignored.

On the base maps,⁶ at each selected point, a slope profile line was extended down the valley wall, orthogonal to the contours, until terminated by either a terrace surface or the present Whitemud channel. By use of the simple trigonometry of a right triangle the average slope was computed. Specifically, this entailed the measurement of the length of the profile line, equivalent to the triangle base (b). This was measured with a micrometer graduated to one hundredths of an inch. The triangle height (c) was estimated to the nearest 5 feet from the contours on the base maps. Calculation of the

⁶Those referred to in footnote 4.

average slope angle (θ) was then given by:

$$\tan \theta = \frac{a}{c}$$

Errors in such calculations were greatest where only a few contours exist; the shallow valley near the Leduc/Calmar highway being the major problem area. Here, delimitation of valley walls was made from aerial photograph inspection but errors of some magnitude are likely. However, in the majority of cases the limits of the valley wall are clearly definable. It is doubtful whether the subjectivity could be greatly reduced for the shallow valley sites even if field measurements were undertaken because breaks in slope must still be arbitrarily assessed (Savigear, 1965).

3:3:3 Whitemud Basin Data

3:3:3:1 Minor Channel and Valley Forms. Certain channel reaches, having the form of grassed-waterways, are essentially indistinguishable from valleys. Such is true for perhaps the greater part of the first order channel lengths. Thus, in carrying out morphometric analysis of channel networks aerial photograph identification of these shallow depressions is the only practicable method. This became clear in the early stages of field work during attempts to measure certain attributes of first order channels on the ground. Many of the channels, clearly visible on the aerial photographs, were impossible to identify with certainty on the ground because of cultivation, crop cover, and simply, indistinct differences between the valley-wall and interfluvial slopes. This was especially true in the southern half of the basin where most first order channels remain

little influenced by the rejuvenation which occurred in the main White-mud valley. Although many valleys, especially in the southern half of the basin, are merely broad linear depressions with almost imperceptibly over-steepened valley-side slopes (Figure 37), these are clearly linked with the basin channel network and contribute concentrated runoff to the channel system under favorable moisture conditions. No attempt was made to fix the downstream limits of such valley forms because the gradation to valleys with incipient floodplains or terraces is very gradual.

Human and animal activities in historic times have been sufficient in the shallow valley areas to make a distinction between natural and artificial forms difficult. To illustrate this point, four channel/valley sites with forms clearly influenced by recent land-use activities were photographed and are briefly described here. Figure 38 illustrates part of a third order channel/valley with grassed-waterway form. However, possibly through weakening of the grass sward by recent over-grazing, the valley has been excavated in places to form plunge-pools similar to those described by Leopold, Wolman, and Miller (1964, p. 450). for the initial stages of discontinuous gully development. A possible explanation is that in seasons of drought the moister valley-bottoms maintained a grass sward longest and, eventually favored spots were overgrazed. During the following spring runoff period these areas could offer little resistance to erosion. Accelerated erosion by cattle-trampling is also very prevalent in the smaller valleys (Figure 39).

Figure 37 - A grassed-waterway near the southern end of the Whitemud basin.

Figure 38 - A grassed-waterway with incipient discontinuous gully development.

Figure 39 - An example of channel/valley form disruption by cattle-trampling.



Recent land-use changes probably account for two well-defined discontinuous gullies (Figures 40 and 41) present in a small third order sub-basin, 6 miles west of Leduc. The two plunge-pools here are in direct contrast to the broad, grassed channel/valley depressions, upstream and downstream, in both depth and steepness of the side-walls and gully-head. Discontinuous gully development appears to be a very local phenomenon within the Whitemud basin, not to be construed as evidence of widespread gullying by headward retreat.⁷ A further example of recent valley-form change is in a reach along the eastern margin of the Leduc golf course where a significant increase in discharge resulted from the construction of large drains by the Leduc overpass on Highway 2. There is a striking difference between the valley upstream of the drain confluence (Figure 42) and that section immediately downstream (Figure 43). Although this study is concerned more with aspects of the natural drainage system of the Whitemud basin, these examples of some unnatural valley modifications show that land use changes, both subtle and obvious, have been operative in modifying channel and valley forms.

3:3:3:2 Terrace Forms. Field surveys revealed that three matched terrace surfaces occupy a considerable length of the main valley, with occasional unpaired remnants at various levels. This

⁷According to the farm owner at one site (Figure 40) this gully-head has retreated over an approximate distance of 25 feet during the last 6 years. A possible contributing factor is that abnormal amounts of water may have been directed into this channel from large drains constructed alongside the Leduc/Calmar highway.

Figure 40 - A discontinuous gully in a small third order basin by the Leduc/Calmar highway.

Figure 41 - Another discontinuous gully in the same third order basin. Note the more stable channel immediately downstream.



Figure 42 - The form of a natural channel/valley reach near the Leduc overpass.

Figure 43 - A drastic alteration of the above channel/valley has been wrought a few yards downstream by the influx of water from a large drain.



was evident from the nearly coincident elevations of many remnant surfaces and from the vertical distribution of the related sediments.

Frye and Leonard (1954) cautioned against use of terrace form only as a means of distinguishing aggradational surfaces. They stressed that the alluvial stratigraphy should be studied as well in the mapping and interpretation of terrace sequences.

Near the Blackmud Creek/Whitemud Creek confluence terrace remnants of the three main surfaces have a vertical distribution such that the upper one lies about 40 feet, the middle terrace between 15 and 20 feet, and the lower terrace surface approximately 10 feet above the present channel. Each terrace consists of approximately 5 to 10 feet of alluvium overlying bedrock or Pleistocene deposits. The contact between bedrock and alluvium is very distinct in that a gravel bed normally forms the basal alluvial unit. The predominant rock types in this gravel are usually characteristic of local till coarse fractions; that is, igneous rocks and dolomites of the Canadian Shield, quartzites from the Rocky Mountains, and local sandstone, shale, and clay-ironstone fragments. There appears to be little systematic change in gravel size with distance up the valley. The material present above the gravel stratum is almost exclusively fine-grained with only occasional lenses of grit and pebbles. At certain sites colluvium from the valley-sides has been mixed with alluvial sediments; at others glacial till and Saskatchewan gravels and sands underlie the Recent fluvial material. Numerically dominant, however, are sections with a stratigraphic sequence of bedrock, alluvial gravel, and fine-grained alluvium. The same sequence generally holds for each of the three terraces.

Figure 44 shows the location of areas mapped in Figures 45 to 48, which in turn, depict the areal distribution of terrace surfaces within that part of the valley where exposures of the constituent sediments occur. The location of each section is indicated on the maps. In addition, Figure 49 illustrates the vertical and horizontal dimensions of terrace remnants within that length of the valley bounded by the Blackmud Creek confluence and the Leduc/Calmar highway.⁸ All terrace remnants between the mouth of the valley and mile 13 (Figure 49) were mapped in the field and checked from the large-scale contour maps. Thirteen miles up the channel the exposures of alluvium become very indistinct. Therefore, the remaining terrace remnants were mapped entirely on the basis of surface elevations (interpreted from the large-scale contour maps). At least the upper two terraces are clearly distinguishable to the Leduc/Calmar highway (Figure 49). The upper terrace appears to continue much further up the main valley from this point, but has not been mapped here.⁹

Generalizations which may be made from Figures 45 to 49 are:

⁸Each terrace remnant as shown in Figure 49 is the hypotenuse of a right triangle. The triangle base approximates the remnant's horizontal dimension; the triangle height denotes the approximate range in elevation of the remnant surface.

⁹Gaps in the mapping of terrace remnants (Figures 45 to 49) occur where coal mining operations have disturbed the valley landforms and where major tributaries enter.



Figure 44 - The Location of Terrace Maps

CONTOURS

DÉPRESSION CONTOURS

INDEFINITE CONTOURS

LAKE RIVER OR STREAM

INTERMITTENT LAKE OR STREAM

DITCH CANAL

SAND OR MUD

POWER TRANSMISSION OR TELEPHONE LINE

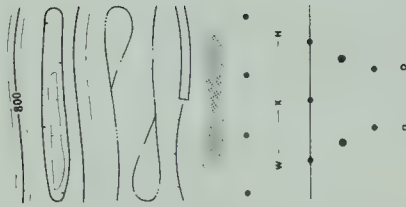
WALL FENCE HEDGEROW

OIL OR GAS PIPELINE

OIL OR GAS TANKS

GROUND CONTROL POINT

MODEL TIE POINT



SWAMP OR MUSKEG

DAM

BRIDGE

GRAVEL PIT OR MINE

WOODED AREA SINGLE TREE

BUILDING - UNDER CONSTRUCTION - EXCAVATION

SPOT ELEVATION

RAILROAD

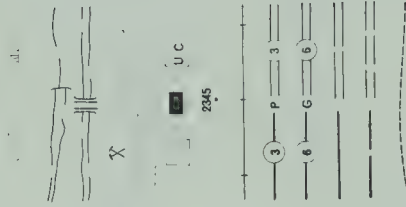
PAVED HIGHWAY

GRAVEL HIGHWAY

WELL TRAVELLED ROAD

NOT WELL TRAVELLED ROAD

TRAIL OR CUT LINE



BASE MAP INDEX

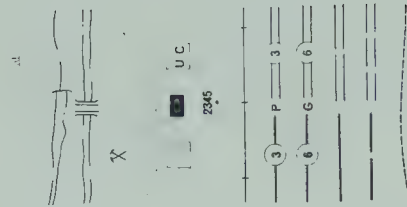


Figure 45 - Terrace Remnants and Section Locations

CONTOURS
 DÉPRESSION CONTOURS
 INDEFINITE CONTOURS
 LAKE RIVER OR STREAM
 INTERMITTENT LAKE OR STREAM
 DITCH CANAL
 SAND OR MUD
 POWER TRANSMISSION OR TELEPHONE LINE
 WALL FENCE HEDGEROW
 OIL OR GAS PIPELINE
 OIL OR GAS TANKS
 GROUND CONTROL POINT
 MODEL TIE POINT



SWAMP OR MUSKEG
 DAM
 BRIDGE
 GRAVEL PIT OR MINE
 WOODED AREA SINGLE TREE
 BUILDING - UNDER CONSTRUCTION - EXCAVATION
 SPOT ELEVATION
 RAILROAD
 PAVED HIGHWAY
 GRAVEL HIGHWAY
 WELL TRAVELLED ROAD
 NOT WELL TRAVELLED ROAD
 TRAIL OR CUT LINE



BASE MAP INDEX

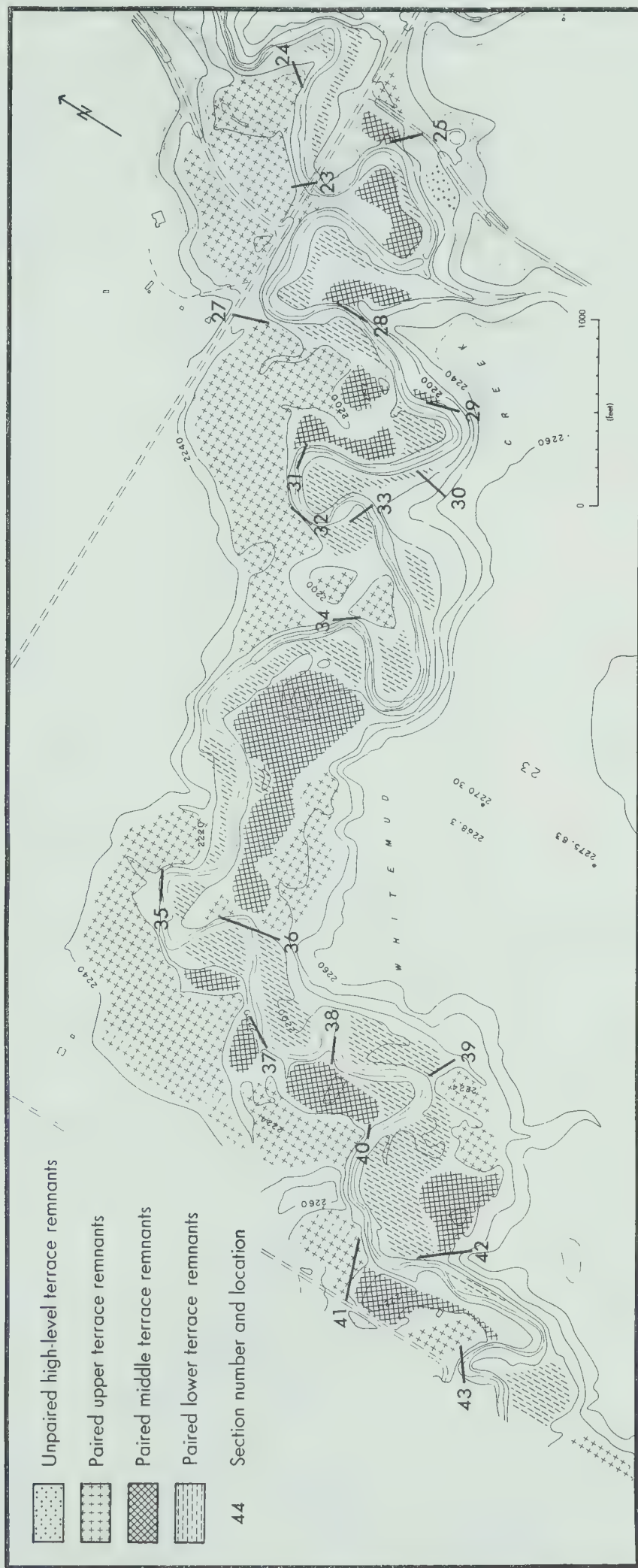
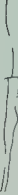
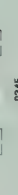
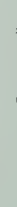
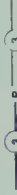
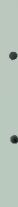


Figure 46 - Terrace Remnants and Section Locations

CONTOURS		SWAMP OR MUSKEG	
DEPRESSION CONTOURS		DAM	
INDEFINITE CONTOURS		BRIDGE	
LAKE, RIVER OR STREAM		GRAVEL PIT OR MINE	
INTERMITTENT LAKE OR STREAM		WOODED AREA, SINGLE TREE	
DITCH, CANAL		BUILDING - UNDER CONSTRUCTION - EXCAVATION	
SAND OR MUD		SPOT ELEVATION	
POWER TRANSMISSION OR TELEPHONE LINE		RAILROAD	
WALL, FENCE, HEDGEROW		PAVED HIGHWAY	
OIL OR GAS PIPELINE		GRAVEL HIGHWAY	
OIL OR GAS TANKS		WELL TRAVELLED ROAD	
GROUND CONTROL POINT		NOT WELL TRAVELLED ROAD	
MODEL TIE POINT		TRAIL OR CUT LINE	

BASE MAP INDEX

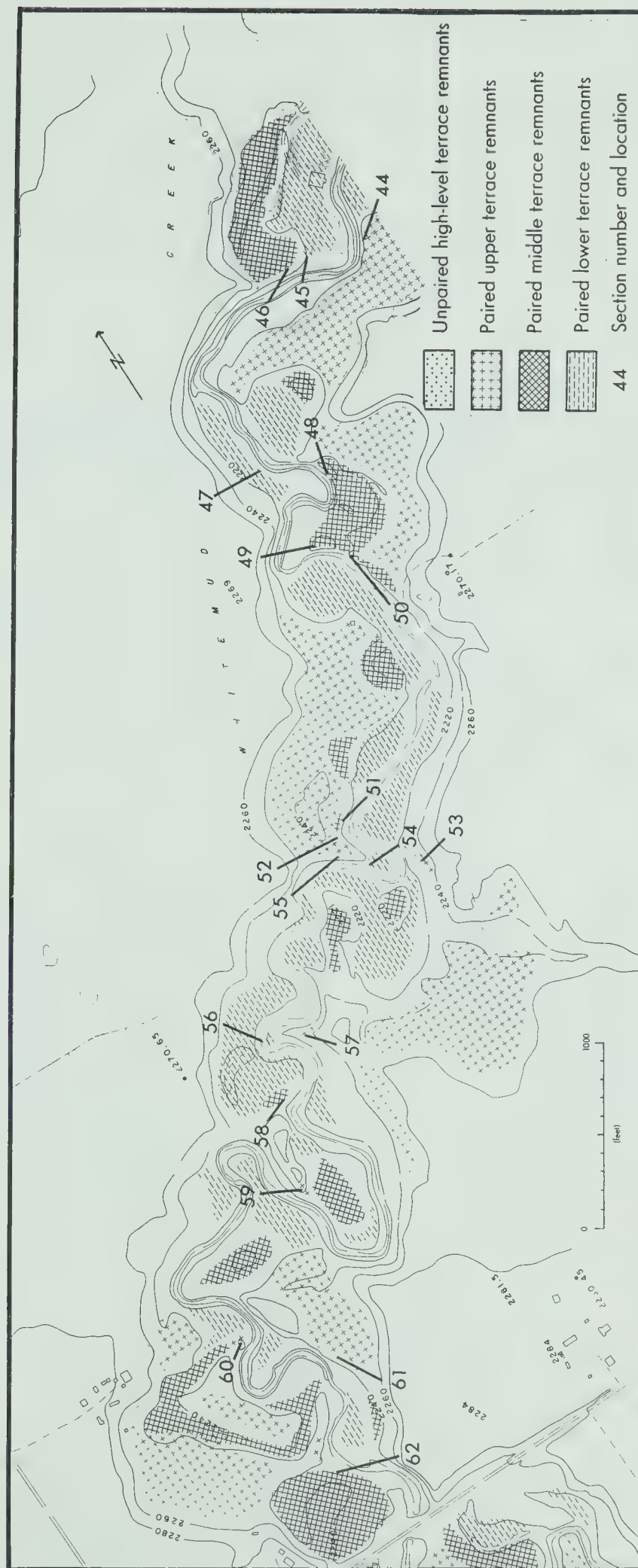
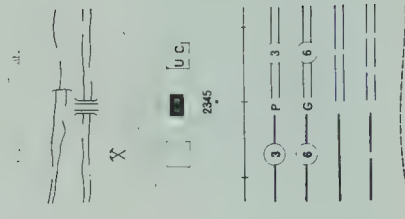
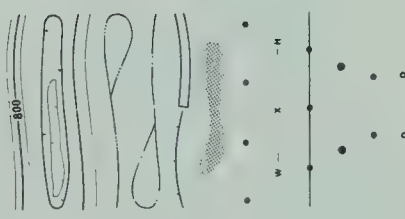


Figure 47 - Terrace Remnants and Section Locations



SWAMP OR MUSKEG
DAM
BRIDGE
GRAVEL PIT OR MINE
WOODED AREA SINGLE TREE
BUILDING - UNDER CONSTRUCTION - EXCAVATION
SPOT ELEVATION
RAILROAD
PAVED HIGHWAY
GRAVEL HIGHWAY
WELL TRAVELLED ROAD
NOT WELL TRAVELLED ROAD
TRAIL OR CUT LINE



CONTOURS
DÉPRESSION CONTOURS
INDEFINITE CONTOURS
LAKE, RIVER OR STREAM
INTERMITTENT LAKE OR STREAM
DITCH, CANAL
SAND OR MUD
POWER TRANSMISSION OR TELEPHONE LINE
WALL FENCE, HEDGEROW
OIL OR GAS PIPELINE
OIL OR GAS TANKS
GROUND CONTROL POINT
MODEL TIE POINT

BASE MAP INDEX

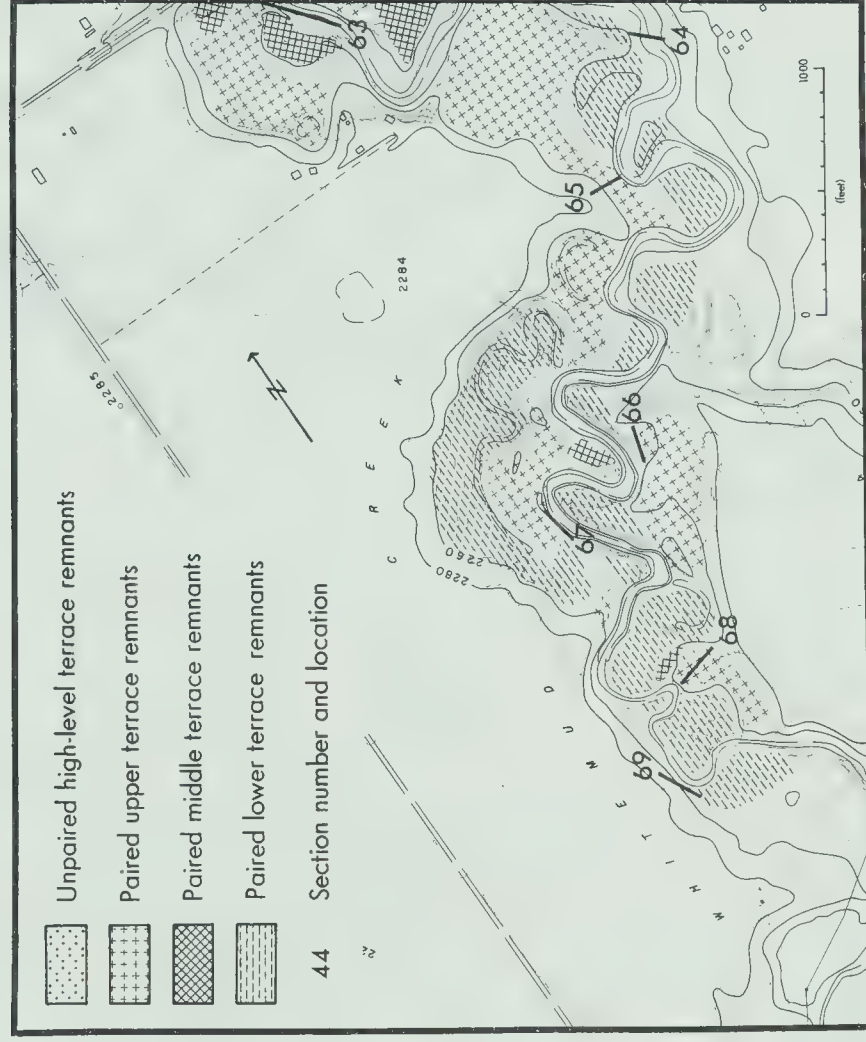


Figure 48 - Terrace Remnants and Section Locations

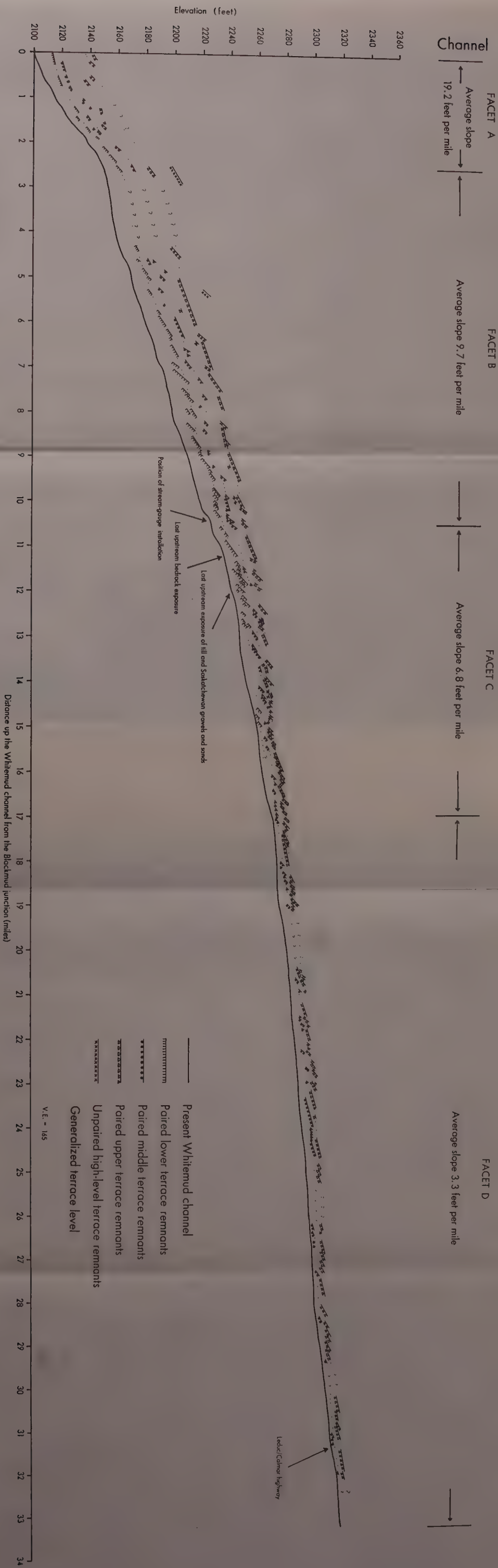


Figure 49 - Partial Longitudinal Profile of Whitemud Creek, and Terrace Remnants

- (1) The upper terrace is marked by only a few small remnants near the downstream end of the valley and at least two of these have been considerably truncated. It is possible that the lower unpaired segments were created during degradation following the upper terrace sediment deposition. Throughout the remainder of the valley the upper terrace is almost ubiquitous.
- (2) The middle terrace remnants are evenly distributed from the Blackmud Creek junction to the Leduc/Calmar highway, at which point they become indistinguishable.
- (3) Remnants of the lower terrace are particularly well-defined in a zone 5 to 13 miles upstream from the valley mouth.
- and(4) Two prominent remnants of a terrace higher than the upper surface previously described occur near the lower end of the Whitemud valley.¹⁰

Variation in the surface elevations of the terrace remnants may be attributed in the main to lateral planation processes during the early stages of degradation following the sediment deposition. At certain sites the bevelled terrace surfaces are marked by successively lower channel scars. The initial aggradational surfaces have also been

¹⁰It should be noted that all terrace remnants (with the exception of the unpaired, high-level ones) have been assigned to one of the three matched terrace levels for purposes of mapping. It is recognized that the remnants which deviate most from the generalized terrace elevations (Figure 49) may well be unmatched surfaces differing in age from the terrace to which they have been linked. This notwithstanding, the three levels are clearly expressed throughout the valley length.

modified by the addition of mass-waste materials from the valley sides. Therefore, the three terrace levels depicted in Figure 49 are based on an approximation of the average elevation of each.

Detailed analysis of the alluvial stratigraphy of the terraces showed that although the three main aggradational sequences exhibit similar gross characteristics subtle differences are present. Following are descriptions of sections selected for mechanical analysis from each of the three terraces. Generalized diagrams (Figures 50 to 52) of all sections studied show the major elements of the depositional sequences. Only the vertical scale is proportional for each diagram and the base-level for each section is the Whitemud channel. Photographs and brief descriptions of other interesting sections are also presented (Figures 53 to 61).

TABLE XIII - THE STRATIGRAPHY OF SELECTED ALLUVIAL SECTIONS

(a) Upper Terrace Section Number	Depth of bed (K)	Description of bed
2	9 inches	SOIL
	2.5 feet	Fine-grained alluvium (sand 35%, silt 38%, clay 27%) with occasional lentoid beds of grit and fine gravel.
	2 feet	Fine-grained alluvium (sand 38%, silt 38%, clay 24%), iron-cemented, with good bedding. Irregular thickness, pinches out in places.
	3 feet	Well-sorted gravel, coarsest near bottom of bed, grading upward to gritty material. Mixture of Shield, Rocky Mountain and local rock types. Predominantly loose matrix, cemented only in patches.

TABLE XIII - THE STRATIGRAPHY OF SELECTED ALLUVIAL SECTIONS

(a) Upper Terrace Section Number	Depth of bed (<)	Description of bed
7	1 foot	Fine-grained alluvium (sand 47%, silt 35%, clay 18%). Iron-cemented. BEDROCK
	35 feet	
	1 foot	SOIL
	5 feet	Relatively homogeneous fine-grained alluvium (sand 42-47%, silt 27-32%, clay 26-28%). Central band of compaction and cementation, up to 1.5 feet deep.
23	2.5 feet	Discontinuous alluvial gravel bed, poorly-sorted with larger fragments of local sandstone, smaller Shield rocks also present. Loose grit and sand matrix.
	35 feet	BEDROCK
	2 inches	SOIL
	2 feet	Artificially deposited detritus.
35	8 feet	Fine-grained alluvium (sand 41-42%, silt 27-30%, clay 29-32%). Good bedding in lower half, moderate in upper portion.
	25 feet	BEDROCK
	1 foot	SOIL
	8 feet	Fine-grained alluvium, upper 3 feet moderately well-bedded (sand 37%, silt 36%, clay 27%), lower 5 feet more compact with good bedding (sand 26%, silt 52%, clay 32%).
53	25 feet	BEDROCK
	9 inches	SOIL
	10 feet	Fine-grained alluvium, more compact in upper 6 feet (sand 62%, silt 21%, clay 17%), relatively loose in the lower 4 feet (sand 74%, silt 12%, clay 14%). Bedding best-developed in lower half.

TABLE XIII - THE STRATIGRAPHY OF SELECTED ALLUVIAL SECTIONS

(a) Upper Terrace		
Section Number	Depth of bed (<)	Description of bed
53	8 feet	Saskatchewan gravels and sands (Rocky Mountain quartzite), well-bedded and sorted, sand matrix.
	10 feet	BEDROCK
(b) Middle Terrace		
Section Number	Depth of bed (<)	Description of bed
4	2 inches	SOIL
	8 feet	Fine-grained alluvium, upper two-thirds sandy (sand 70-74%, silt 10-12%, clay 16-18%), lower one-third less so (sand 52%, silt 27%, clay 21%). Well-bedded and relatively loose throughout.
	8 feet	Gravel alluvium, mixture of Shield, Rocky Mountain, and local rocks. Loose grit and sand matrix. Full depth of bed not visible.
9	5 inches	SOIL
	2 feet	Fine-grained alluvium (sand 33%, silt 36%, clay 31%) with good bedding. Relatively non-compacted.
	8 feet	Fine-grained alluvium dominated by silt (sand 17-26%, silt 49%, clay 26-34%). Compact and well-bedded with occasional lentoid beds of grit and pebbles. Contains a spruce tree stump, apparently in the position it grew, which is truncated at the contact with the overlying alluvial stratum.
	1 foot	Poorly exposed alluvial gravel composed of Shield, Rocky Mountain and local rocks.
	4 feet	BEDROCK
37	8 inches	SOIL
	6 feet	Fine-grained, homogeneous alluvium (sand 45-46%, silt 31-33%, clay 22-33%) with moderate

TABLE XIII - THE STRATIGRAPHY OF SELECTED ALLUVIAL SECTIONS

(b) Middle Terrace Section Number	Depth of bed (<)	Description of bed
37	1.5 feet	bedding in the upper half, good bedding in the lower half. Fine-grained alluvium consisting mainly of pebbles and grit. Dominated by local shale, with a secondary element of Shield and Rocky Mountain rocks.
	1 foot	Coarser gravel alluvium with local clay-ironstone predominant. Shield and Rocky Mountain rocks present also, with comminuted local shale forming the bulk of the matrix.
	6 feet	Good bedding. BEDROCK
51	6 inches	SOIL
	5 feet	Fine-grained, relatively homogeneous alluvium (sand 38-45%, silt 30-38%, clay 23-25%). Moderately well-bedded throughout but developed best toward bottom. Compaction also increases towards bottom.
	2 feet	Coarse gravel alluvium composed of Shield, Rocky Mountain and local rocks. Moderately well-bedded with loose grit and sand matrix.
	8 feet	BEDROCK
(c) Lower Terrace Section Number	Depth of bed (<)	Description of bed
1	1 inch	SOIL
	3 feet	Colluvium.
	7 feet	Fine-grained homogeneous alluvium, (sand 63-66%, silt 19%, clay 15-18%). Poorly-bedded, loose material. Full depth of bed not visible.

TABLE XIII - THE STRATIGRAPHY OF SELECTED ALLUVIAL SECTIONS

(c) Lower Terrace Section Number	Depth of bed (←)	Description of bed
16	3 inches	SOIL
	6 feet	Fine-grained alluvium, sand dominant in upper half (sand 76%, silt 12%, clay 12%) but less so in lower half (sand 47%, silt 30%, clay 23%). Moderately well-bedded throughout, with a central layer of relatively compacted sediment.
	3 feet	Poorly-exposed alluvial gravel of Shield, Rocky Mountain and local varieties. Loose grit and sand matrix. Full depth of bed not visible.
19	4 inches	SOIL
	5 feet	Fine-grained alluvium, sand dominant in the upper half (sand 67%, silt 15%, clay 18%) less so in the lower portion (sand 46%, silt 28%, clay 26%). Good bedding throughout.
	2 feet	Coarse gravel alluvium consisting of Shield, Rocky Mountains and local rocks. Poorly-sorted and bedded but tends to coarsen towards the bottom. Loose grit and sand matrix.
3 feet	BEDROCK	
30	6 inches	SOIL
	1 foot	Fine-grained alluvium (sand 71%, silt 13%, clay 16%), loose and well-bedded.
	4 feet	Fine-grained alluvium (sand 51%, silt 27%, clay 22%), loose, and moderately well-bedded.
	1 foot	Fine-grained alluvium (sand 31%, silt 47%, clay 22%), moderately well bedded.
	3 feet	Fine-grained alluvium (sand 54%, silt 27%, clay 19%), loose, and moderately well-bedded.
3 feet	BEDROCK	

TABLE XIII - THE STRATIGRAPHY OF SELECTED ALLUVIAL SECTIONS

(c) Lower Terrace Section Number	Depth of bed (<)	Description of bed
39	4 inches	SOIL
	2 feet	Fine-grained alluvium (sand 17%, silt 47%, clay 36%), relatively loose with moderate bedding.
	6 feet	Fine-grained alluvium (sand 47-55%, silt 26-30%, clay 19-23%), loose and moderately well-bedded. Full depth of bed not visible.
54	4 inches	SOIL
	3 feet	Fine-grained alluvium, loose in upper half (sand 73%, silt 13%, clay 14%), moderately compacted in lower portions (sand 62%, silt 19%, clay 19%). Good bedding throughout.
	3 feet	Gravel alluvium, fine and well-sorted, with pronounced imbrication. Dominated by local shale but with incorporated Shield and Rocky Mountain rocks. Full depth of bed not visible.

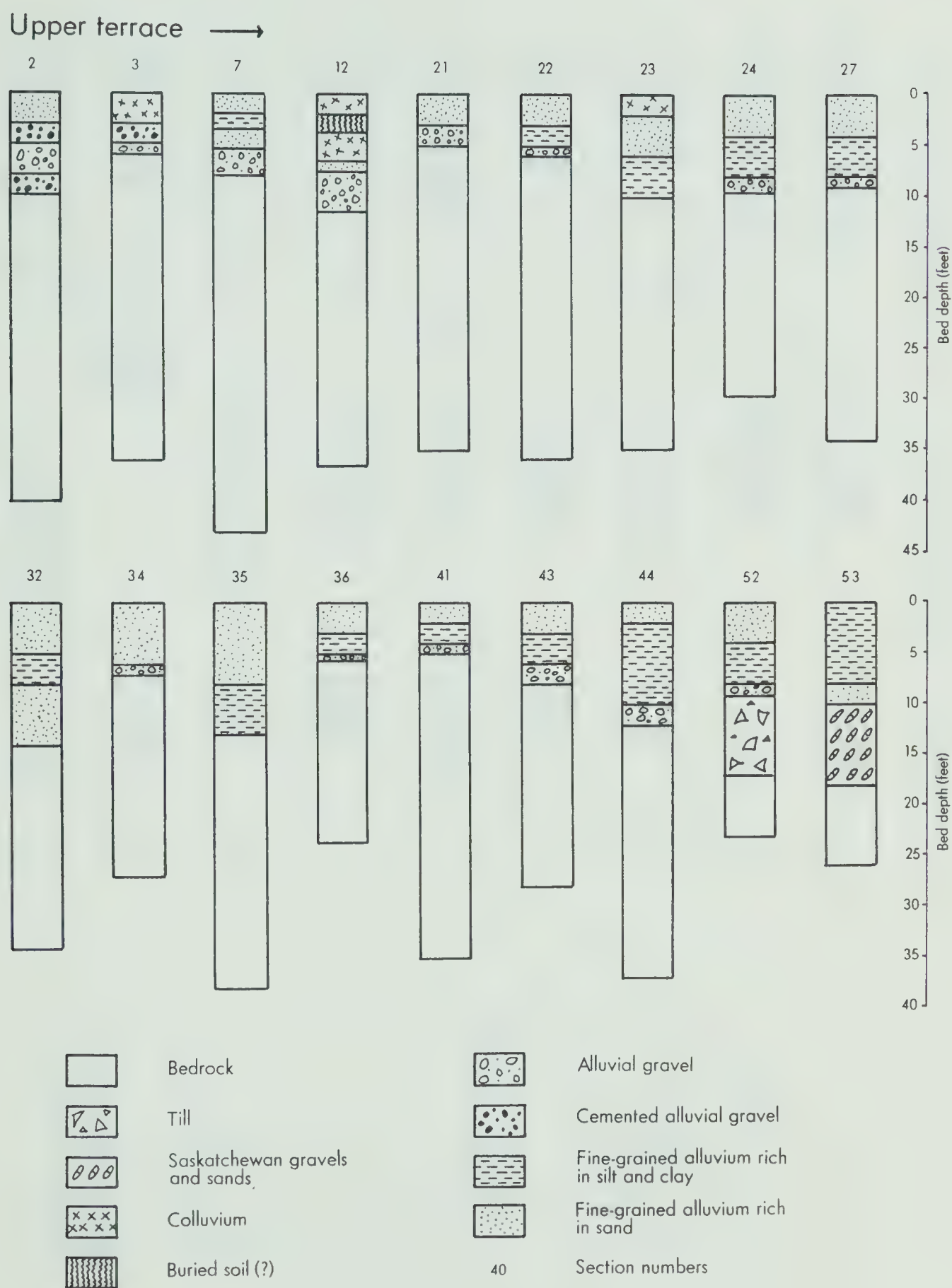


Figure 50 - Generalized Sections of Terrace Alluvium

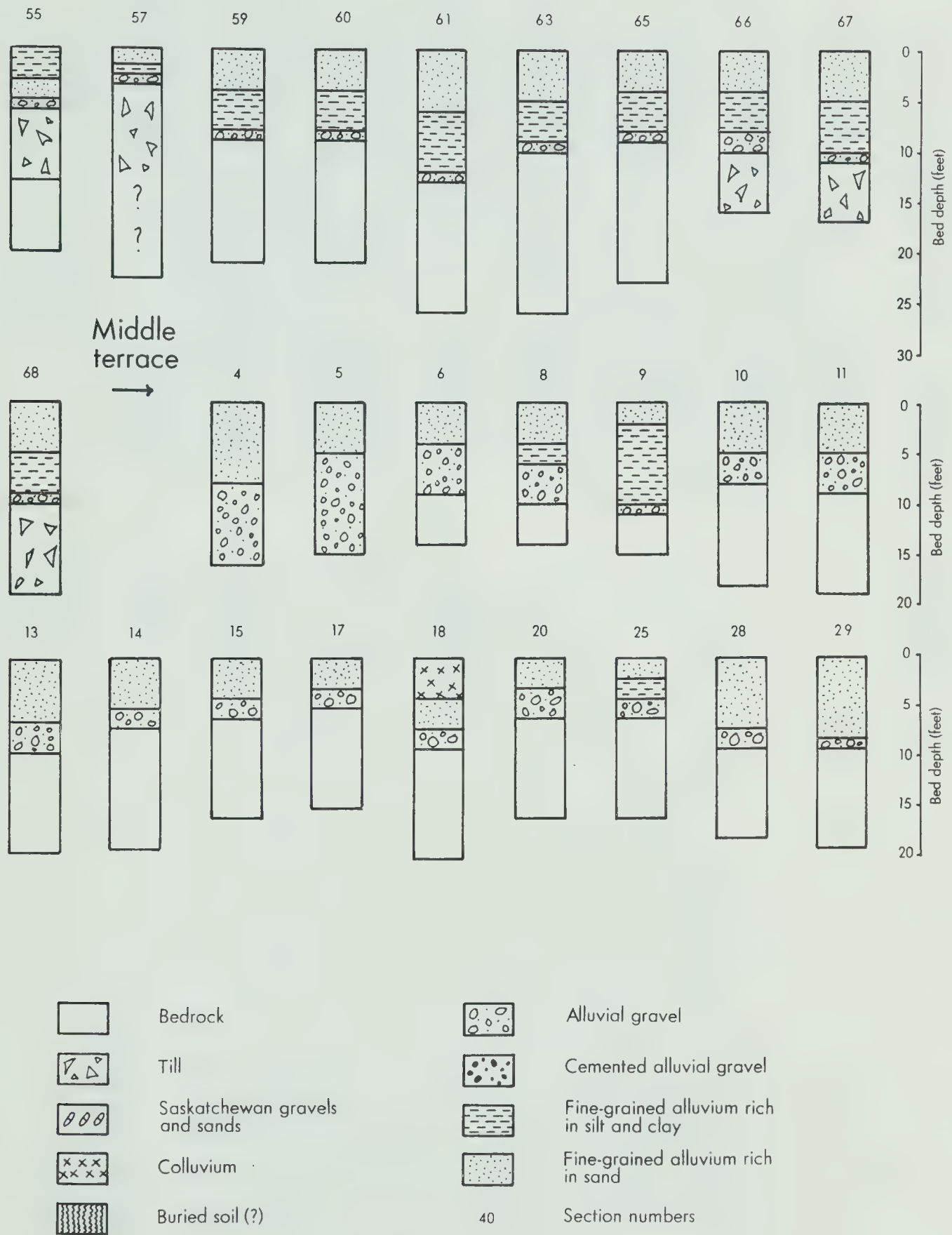


Figure 51 - Generalized Sections of Terrace Alluvium

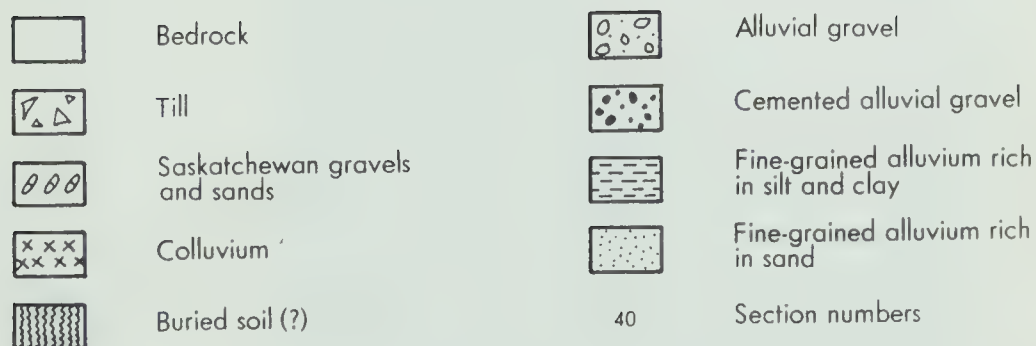
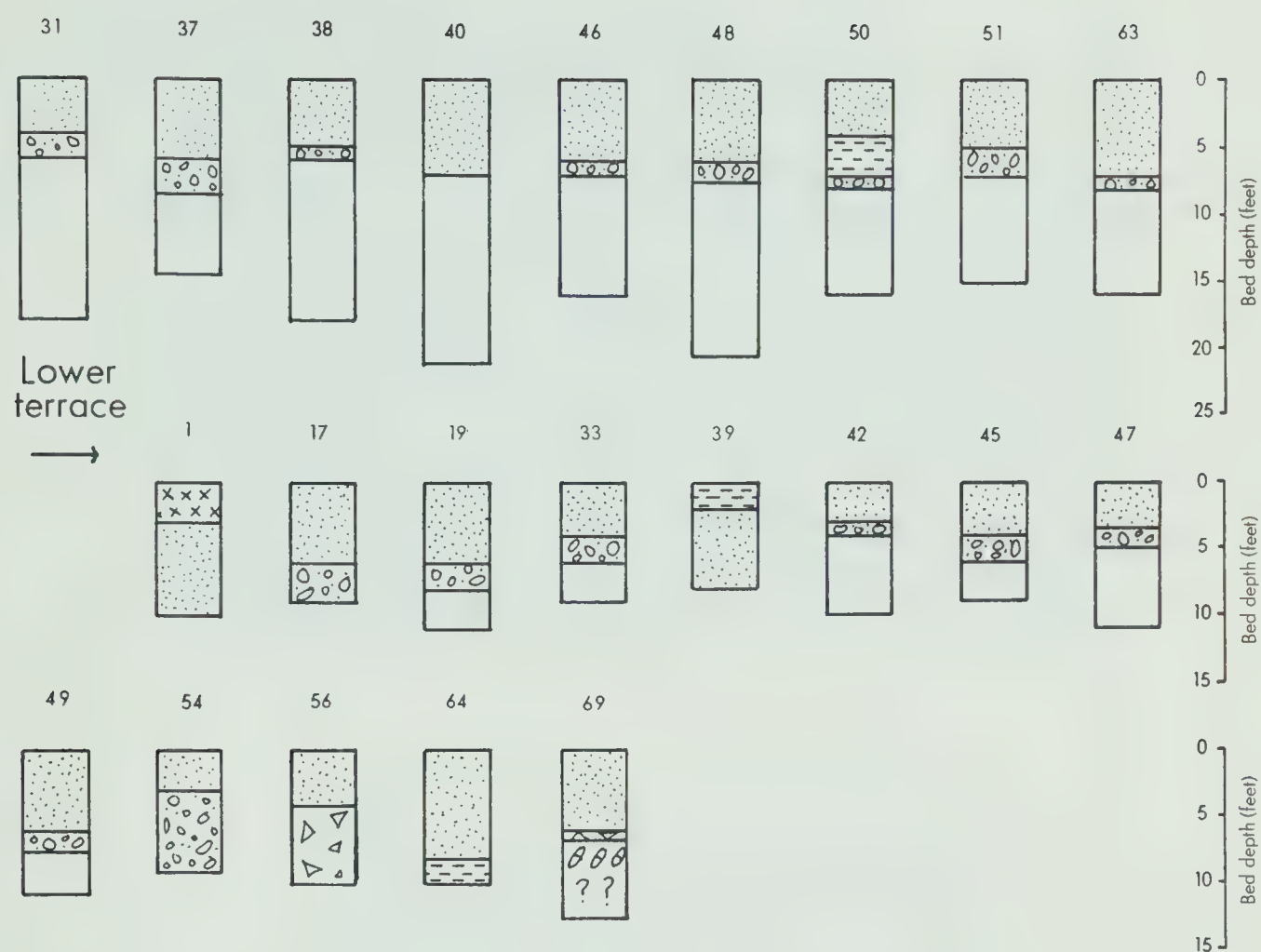


Figure 52 - Generalized Sections of Terrace Alluvium

Figure 53 - The upper terrace alluvium shown here is that of Section 21. The shale bedrock is succeeded by 2 feet of relatively well-bedded alluvial gravel, composed mainly of Shield and Rocky Mountain rocks. This material coarsens markedly towards the bottom where slabs of local shale occur. Well-bedded, fine-grained alluvium, 3 feet deep, overlies the gravel. The related terrace surface stands approximately 30 feet above an abandoned channel cut into the lower terrace.

Figure 54 - Well-illustrated in Section 22 is the threefold character of alluvium observable at many sites. Immediately above the shale and sandstone bedrock a bed, 1 foot thick, of gravel occurs. Succeeding this is approximately 2 feet of very compact, well-bedded alluvium which is comparatively rich in silt and clay. In turn, this stratum is overlain by loose, sandy sediments with moderately good bedding. The present Whitemud channel passes by the base of the exposure.

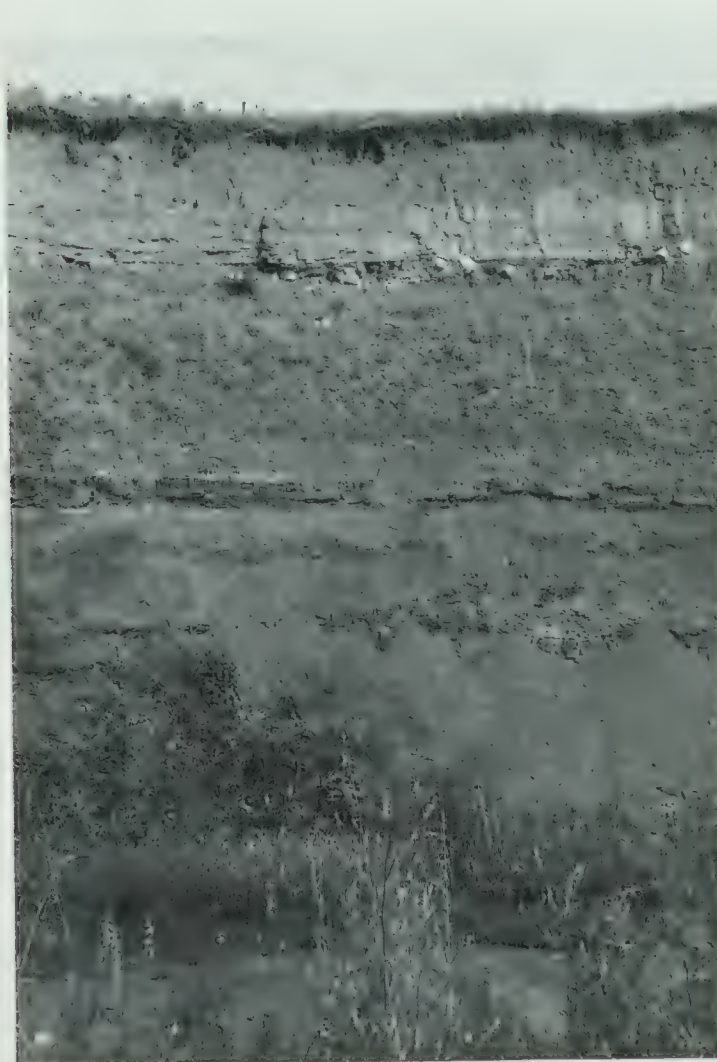


Figure 55 - The site of Section 55 is unusual in that a remnant of the upper terrace is flanked by a portion of the lower one. Underlying 5 feet of moderately well-bedded, fine-grained alluvium and 1 foot of alluvial gravel, in the upper terrace, is a bed of glacial till 7 feet thick. Beneath the till bedrock is clearly visible. What appears to be a buried soil occurs under part of the fine-grained alluvium forming the flanking, lower terrace (to the right of the geology hammer in Figure 55a). Major subdivisions of the material in this exposure are depicted diagrammatically in Figure 55b.

(a)



(b)

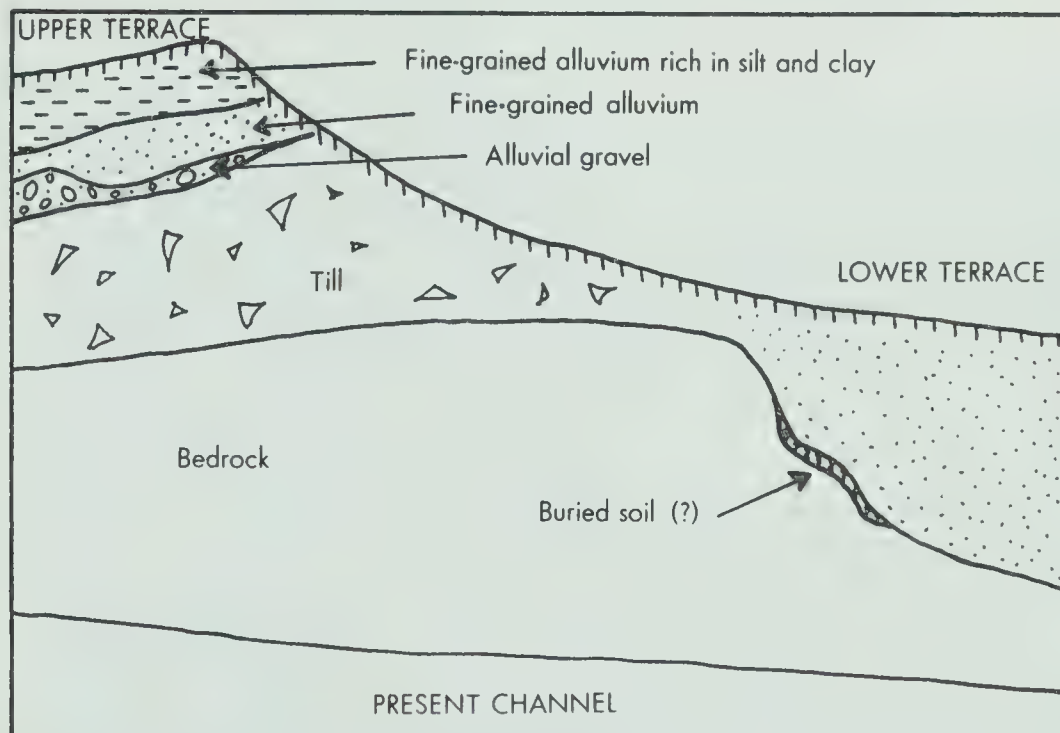


Figure 56 - Section 13 is an example of the middle terrace stratigraphy. Ten feet of bedrock rises above the present Whitemud channel and is immediately overlain by up to 3 feet of well-sorted gravel. This gravel is coarsest near the bedrock contact and grades into relatively fine material above. Shield, Rocky Mountain and local rock types occur within the bed. Poor to moderately well-bedded, fine-grained alluvium, to a depth of 6 feet in places, overlies the gravel. This exposure continues for 75 yards along the present channel, with little variation from the sequence described here.

Figure 57 - Section 29, an exposure of the middle terrace material, is distinctive in that it is flanked immediately down-valley by a remnant of the lower terrace. The middle terrace stratigraphy is almost identical to that of Section 13 except that the gravel bed is only 1 foot thick and the finer sediments attain a depth of 8 feet.



Figures 58 to 60 - At one site (Section 68) the three terrace surfaces are exposed side by side, where the present Whitemud channel has truncated these deposits. Figure 58 (top) illustrates the middle terrace. The coarse gravel in the foreground is a recent deposit unrelated to the terrace. This material obscures the lower strata of the exposure but moderately well-bedded, fine-grained material is clearly visible. To the right may be seen part of the terrace face which rises to the upper terrace surface, and the alluvium underlying the latter is shown in Figure 59 (middle). Approximately halfway between the terrace surface and the present channel a thin bed of gravel overlies glacial till. Well-bedded, fine-grained sediments, up to 9 feet deep, rest on the gravel. A few yards upstream a steep terrace face leads from the upper surface to the lower terrace. The alluvium of the lowest terrace is poorly-exposed but the related shallow channel-scar is visible in Figure 60 (bottom).

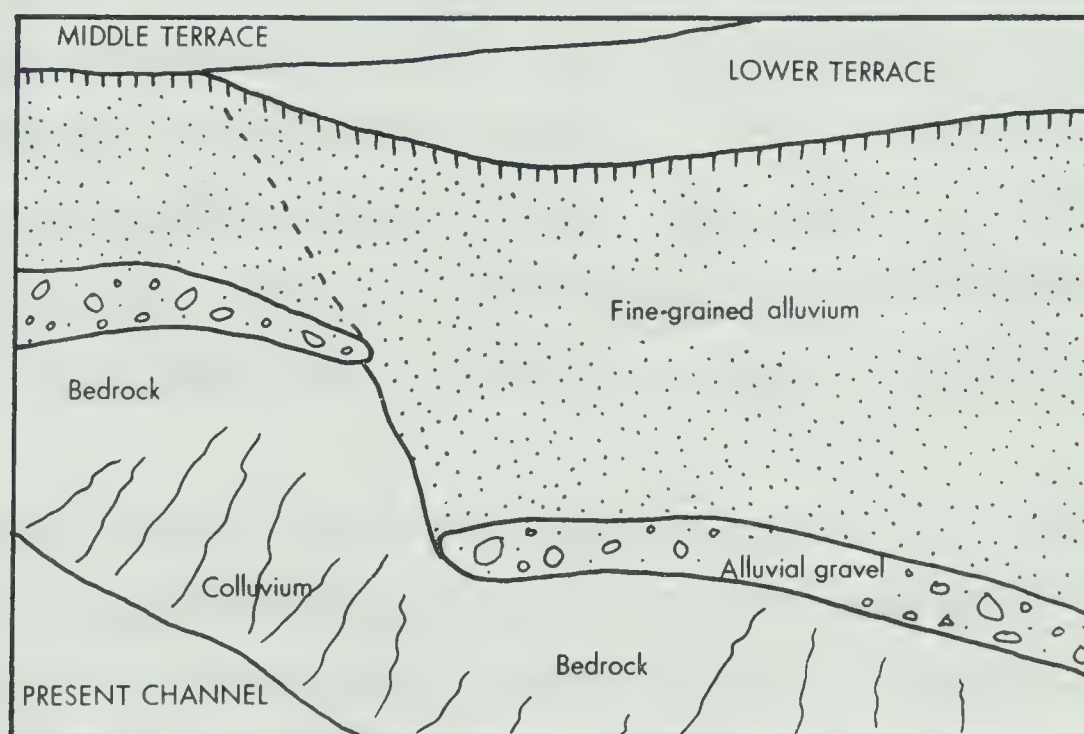


Figure 61 - The relationship between the middle and lower terraces is clearly illustrated at the site of Section 45. Here the bedrock, gravel, fine-grained sediment, sequence is evident for both terraces as is the vertical displacement of one above the other. Separating the two surfaces is a low-angle terrace face. In each case bedrock is mantled by about 2 feet of gravel and 4 feet of finer material (Figure 61a). Local clay-ironstone dominates the gravel strata. Figure 61b shows more clearly the main features of this site.

(a)



(b)



Taking each terrace in turn we may note first, from Figures 50 to 52, that of twenty-eight sections for the upper unit, twenty-two have a gravel layer immediately above the bedrock or Pleistocene material. Eighteen have a well-bedded, relatively compact, silt/clay-rich bed above the gravel and twenty-two exhibit fairly loose, sandy sediments in their upper portions. Second, the middle terrace has a somewhat comparable alluvial sequence but here the bedding, compaction and silt/clay content in the lower part of the fine-grained material seems to be less pronounced than for the higher terrace. Thus, of twenty-five sections, twenty-four have gravel at the base of the alluvium, three have silt/clay-rich beds, similar to those of the upper terrace, overlying the gravel and all have relatively loose, sandy material forming the bulk of the sequence. Third, the lower terrace has only thirteen good sections, of which eight contain gravel at the alluvial base. All sections are dominated by loose, sandy sediments with only slight improvement of bedding and compaction towards the bottom.

A more precise illustration of the differences in the fine-grained alluvium of each terrace is given by the frequency distributions of values determined by grain-size analysis (Figure 62). It would appear that there may be little difference between the average textures of the upper and middle terraces, even though the higher unit has the better-defined silt and clay-rich stratum. For small sample numbers, as in this case, it is appropriate to use the Student t test to ascertain whether significant differences between the means actually exist. In applying this test the following

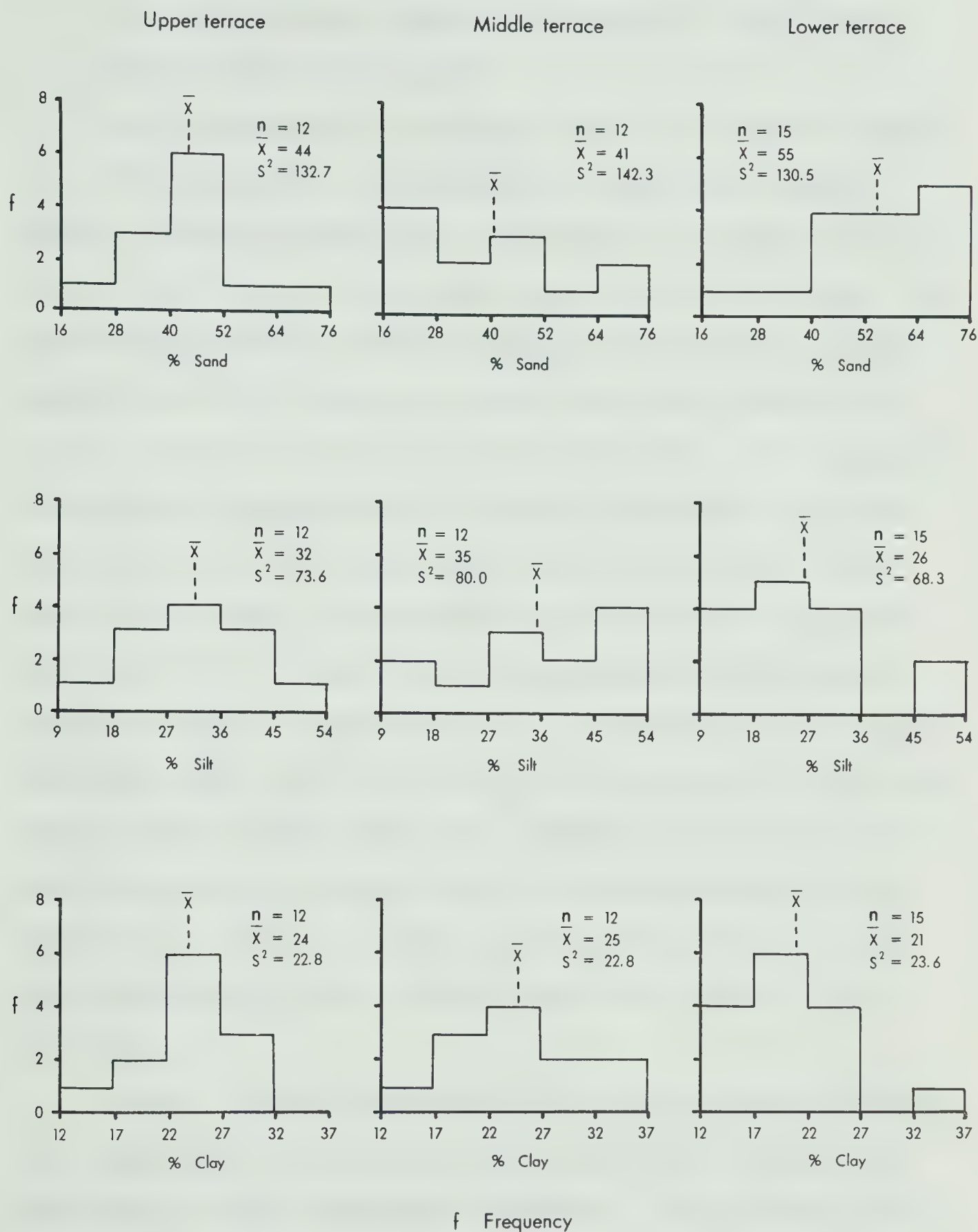


Figure 62 - Sand, Silt and Clay Percentages
for Terrace Alluvium Samples

assumptions must be made (Strahler, 1954):

- (1) the populations sampled are independent of each other
- (2) the sampling was random
- (3) the population distributions must be approximately normal
- and (4) the population variances must be approximately equal.

Clearly, the three terraces may be considered as independent units, insofar as their distribution indicates they are of different age. Although three of the nine sample frequency distributions are markedly skewed (Figure 62), it seems that most approximate a normal distribution. Therefore, on the basis of this sample data, it is reasonable to assume that the populations are normally distributed. The homogeneity of population variances may be verified using the F test (Cole and King, 1968). This assumption was satisfied in the present case (Table XIV). The sampling technique employed cannot be considered strictly as a random method but, nevertheless, it was thought reasonable to apply the test in that any bias during collection of the material would have been unconscious. However, because of this limitation the conclusions reached from this test application should be considered as tentative. Further studies of the Whitemud alluvium using a more rigorous sampling method may provide more definitive conclusions.

Briefly, the test procedure was as follows: it was postulated (null hypothesis, H_0) that mean values of sand, silt, and clay for each terrace are not significantly different. Thus, each mean value of percentage sand was tested against the other, and so on, for the

TABLE XIV - RESULTS OF TESTS FOR HOMOGENEITY OF VARIANCES AND DIFFERENCES OF MEANS OF TERRACE ALLUVIUM TEXTURES

F test Grain size	Units Tested	F_c	F	Test Result*
SAND	a_1b_1	2.82	1.07	Accept H_0
	a_1c_1	2.56	1.02	Accept H_0
	b_1c_1	2.56	1.09	Accept H_0
SILT	a_2b_2	2.82	1.09	Accept H_0
	a_2c_2	2.56	1.08	Accept H_0
	b_2c_2	2.56	1.17	Accept H_0
CLAY	a_3b_3	(identical variances)		Accept H_0
	a_3c_3	2.74	1.04	Accept H_0
	b_3c_3	2.74	1.04	Accept H_0
t test Grain size	Units Tested	t_c	t	Test Result*
SAND	a_1b_1	2.07	0.63	Accept H_0
	a_1c_1	2.06	-2.47	Reject H_0
	b_1c_1	2.06	-3.10	Reject H_0
SILT	a_2b_2	2.07	0.84	Accept H_0
	a_2c_2	2.06	1.84	Accept H_0
	b_2c_2	2.06	2.71	Reject H_0
CLAY	a_3b_3	2.07	0.51	Accept H_0
	a_3c_3	2.07	1.59	Accept H_0
	b_3c_3	2.07	2.13	Reject H_0

*Tested at the 5% level of significance

three size ranges. Table XIV outlines the test results, at the 5 per cent significance level and with degrees of freedom as calculated in the Student t test equation (Appendix C). Here the upper terrace sand mean is designated by the symbol, a_1 , the silt mean a_2 , and the clay mean a_3 . The two lower terraces are denoted by letters b and c, with similar subscripts.

In sum, the difference of means test indicated that the upper and middle terrace have average sand contents significantly different from the lower terrace. By the same token, there is no significant difference between the mean sand values of the upper and middle units. Further, the middle and lower terraces differ in their silt and clay content. All other combinations within each grain-size class have comparable mean values. It should be reiterated that these conclusions, while apparently reasonable, are tentative because of the non-random sampling method employed.

Plotting of the average values on a texture diagram (Figure 63)¹¹ revealed that the upper and middle terraces are composed of alluvium in the loam class while the lower terrace is transitory between loam and sandy loam. It seems probable that the three terraces have been

¹¹The mechanical grain-size analysis for this work was based on the size classification of the U.S.A. Department of Agriculture, as is the texture diagram (Figure 63).

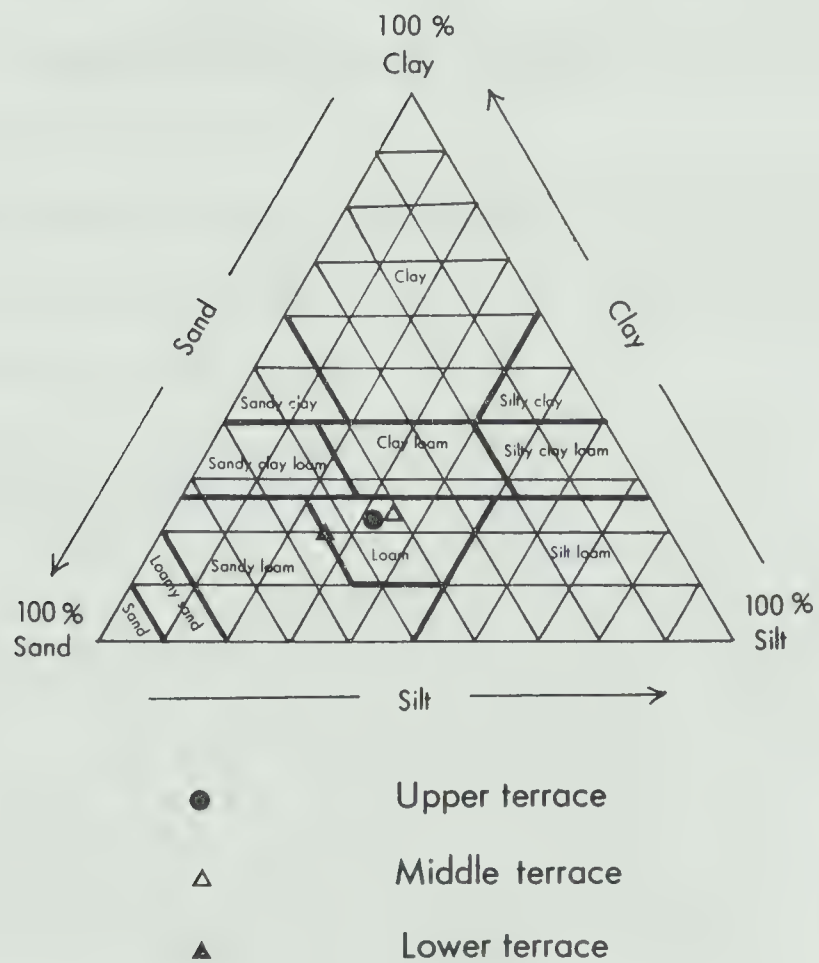


Figure 63 - Average Textures of the Fine-grained Terrace Alluvium

formed by comparable processes. Gross similarities in the depositional sequences are quite evident.

It has been suggested (see: Schumm and Lichty, 1963; King, 1966; Morisawa, 1968) that deposits of a relatively coarse nature may often be attributed to lateral accretion during channel shifting. Fine-grained sediments tend to accumulate by vertical accretion during and immediately following flood-peak discharges.

Two recent papers (Allen, 1963; Visher, 1965) contain valuable summaries of fluvial processes, sedimentary structure/texture relationships. Allen (1963, p. 166) stated:

It has long been understood in a general way that a river flowing through alluvium deposits under distinct circumstances: within the confines of the channel at ordinary river stages, and on the floodplain top-stratum, in addition to within the channel, at high or flood stages (Wolman and Leopold, 1957). The deposited stream load then becomes differentiated texturally. The relatively coarse materials carried on or close to the bed — the gravels and coarser sands — remain within the channel, while the relatively fine materials maintained in suspension by the turbulent motion of the water — the finer sands and silts and clays — are deposited beyond the channel banks on the floodplain top-stratum.

Visher (1965, p. 132) presented a fluvial model representing an idealization of the vertical distribution of grain-size and sedimentary structures on a point bar. The ideal texture sequence, based on a consideration of flow regime conditions, is in ascending order, medium-coarse grains, fine-medium grains, and very fine grains. The related process elements are in general terms, successively, lag and lateral accretion deposition, current lamination, and overbank deposition.

The vertical texture distribution of sediments in the middle

and upper terraces of the Whitemud valley appear to deviate substantially from this model sequence in that the very fine deposits immediately succeed the lowermost coarse gravel stratum. In turn, the silt and clay-rich bed is overlain by slightly coarser, sandy sediments. Furthermore, the lower terrace has relatively homogeneous sandy alluvium above the gravel, with little evidence of an upper silty and clayey layer.

3:3:3:3 Valley-wall Slopes. The frequency distributions and descriptive statistics derived for the east and west valley-wall slopes indicate that asymmetry is very unlikely (Figure 64). Both frequency distributions are slightly right-skewed but plotting on arithmetic and logarithmic probability paper (Chorley, 1966) showed that they lie somewhere between normal and log-normal forms. The skewness may be attributed to a small number of very steep slopes produced by basal sapping near the basin mouth. However, since valley-wall mean slope angles usually have approximately normal distributions (Carter and Chorley, 1961; Krumbein and Graybill, 1965) it was assumed for purposes of analysis that this is true for the Whitemud populations. It was not necessary to test the means and variances of the total east and west wall sample distributions because they are clearly similar (Figure 64). Therefore, the analysis of mean slopes above various surfaces could reasonably be carried out using composites of sample values from both populations. Application of the F test for homogeneity of variances indicated that the Student t test could be applied to sample means of:

- (1) the present channel and upper terrace groups
- (2) the middle and lower terrace groups

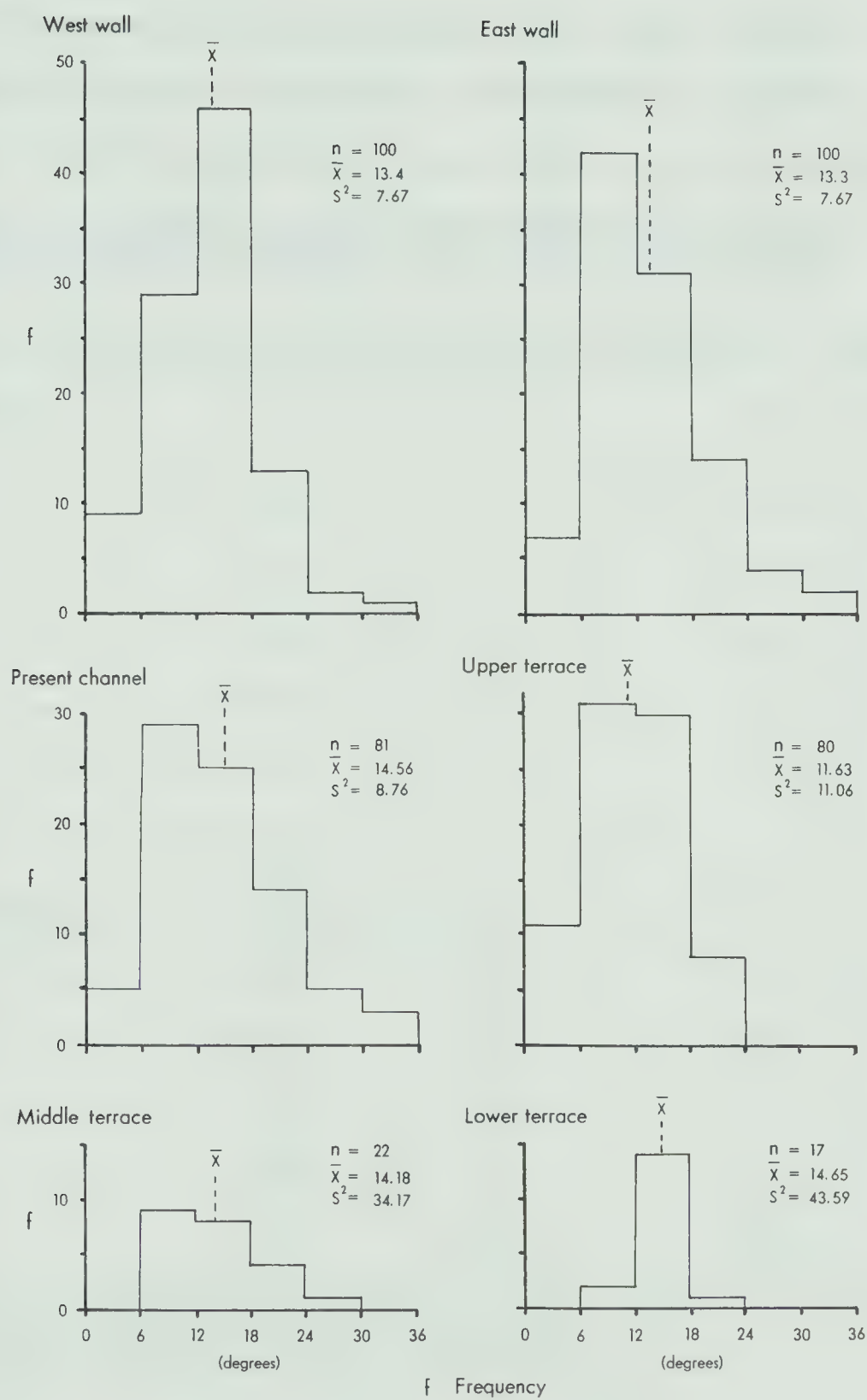


Figure 64 - Valley-wall Slopes

and (3) paired combinations of consecutive up-valley segments.

Although the mean values for slopes terminating, for example, at the present channel and lower terrace limits, are nearly identical (Figure 64) they could not be analyzed by the difference of means test used here. Table XV contains the F and t test results obtained for

TABLE XV - RESULTS OF TESTS FOR HOMOGENEITY OF VARIANCES AND DIFFERENCES OF MEANS OF VALLEY-WALL SLOPE SAMPLES

F Test Samples tested	F_c	F	Test Result*
P.C./U.T.	1.44	1.26	Accept H_0
M.T./L.T.	2.15	1.28	Accept H_0
A/B	1.61	1.12	Accept H_0
B/C	1.61	1.35	Accept H_0
C/D	1.61	1.07	Accept H_0
t test Samples tested	t_c	t	Test Result*
P.C./U.T.	1.96	5.88	Reject H_0
M.T./L.T.	2.02	-0.24	Accept H_0
A/B	1.99	4.21	Reject H_0
B/C	1.99	4.50	Reject H_0
C/D	1.99	-0.83	Accept H_0

*Tested at the 5% level of significance.

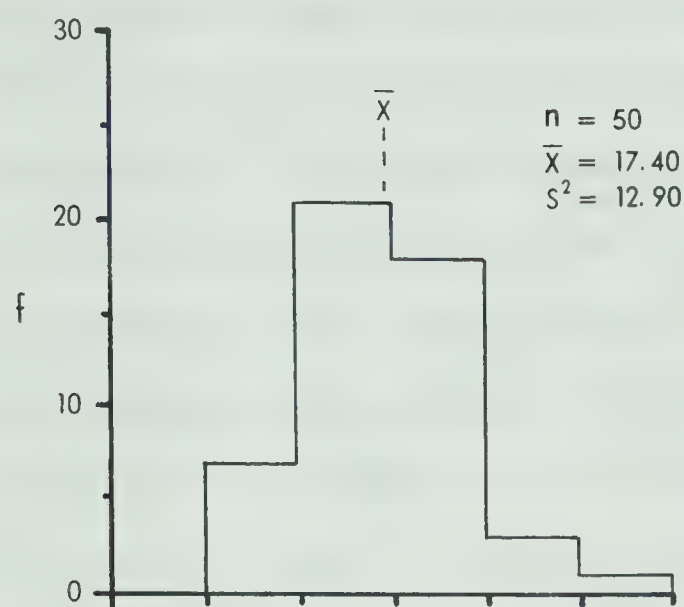
those combinations which were tested. The notation used to distinguish particular sample sets is as follows: present channel (P.C.), upper terrace (U.T.), middle terrace (M.T.), lower terrace (L.T.), and con-

secutive segments upstream from the Blackmud Creek junction, A, B, C and D, respectively.

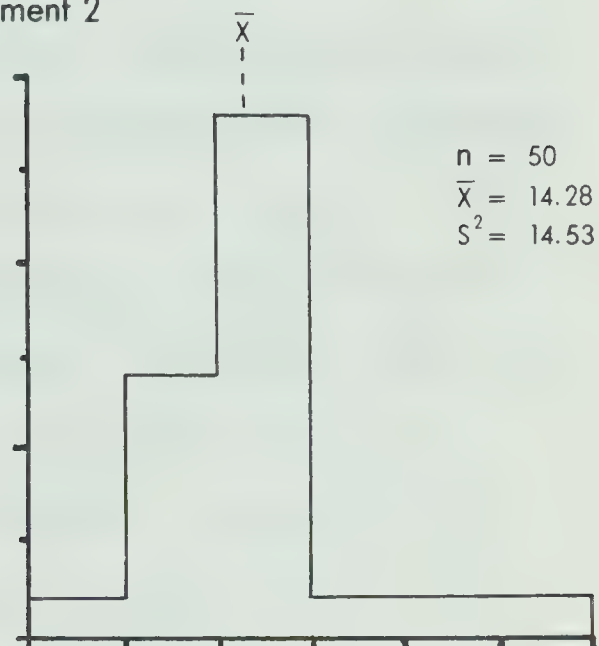
On this basis it is suggested that significant differences exist between valley-wall slope angles of the three valley segments closest to the basin mouth. The fourth segment, on the other hand, has slope angles differing insignificantly from those of the adjacent downstream segment. However, the magnitude of measurement-error is sufficient for this part of the valley to preclude definite conclusions. Figure 65 illustrates the change in mean valley-wall slope angles from segment to segment. Near the basin mouth these slopes are steepest, with a decline of approximately 3 degrees between each of the other segments. The influence of lithologic variation on slopes here is a topic requiring detailed field investigation, outside the scope of this study. However, casual field observations indicate that marked slope differences occur only where either clay-ironstone or relatively resistant sandstone strata are present. For the most part, the local bedrock has a high bentonite content and is thus incapable of maintaining slopes steeper than those of the overlying Pleistocene deposits. It is difficult, therefore, to determine the bedrock-glacial deposit contact on the basis of slope form alone.

The evidence indicates further that the oldest portions of the valley walls, those above the upper terrace surface, may have declined in the order of 3 degrees to a present mean angle of about 11.5 degrees (Figure 64). This inference rests on the assumption that an average slope angle of approximately 14.5 degrees is characteristic of the more recently formed slopes. These are, of course, the slopes termi-

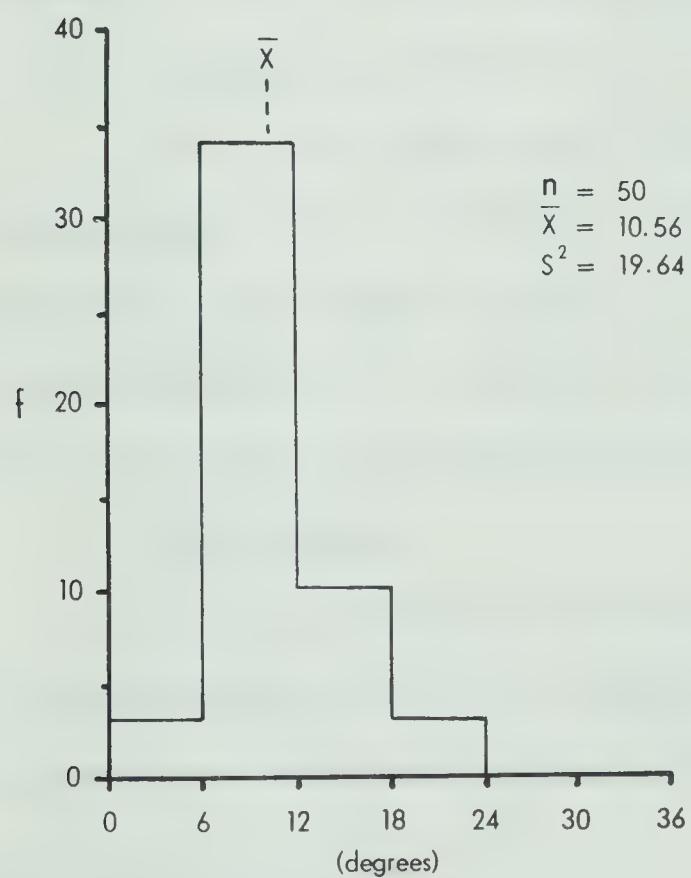
Segment 1



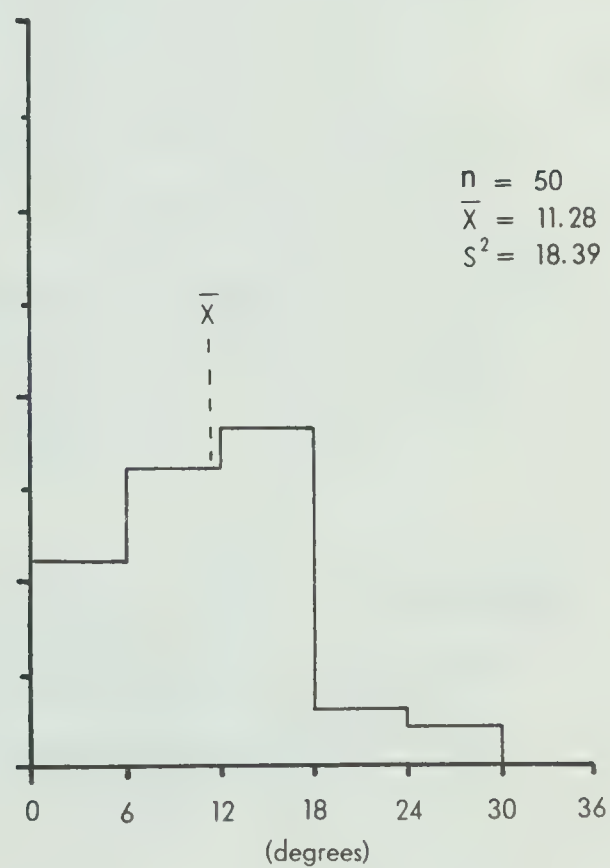
Segment 2



Segment 3



Segment 4



f Frequency

Figure 65 - Valley-wall Slopes

nating at the present channel, the lower terrace and middle terrace (Figure 64). Variations in sampling might have produced the difference of 3 degrees between the average angle of slopes bordered by the upper terrace (oldest) and those terminating at the present channel (youngest). For example, if the upper terrace remnants are closely spaced in any given valley segment the probability of acquiring a disproportionately large sample of the oldest slopes is increased. As a check, therefore, the frequency distributions of the oldest and youngest slopes were plotted for each segment of the valley (Figure 66). Regardless of differences in sample sizes the mean angles of slopes leading to the present channel are consistently steeper than those above the upper terrace.¹² This reinforces the suggestion that the older slopes have in fact declined significantly since they were cut initially.

In sum, the three tentative hypotheses regarding valley-wall slopes appear to be justifiable; there is no evidence of valley asymmetry, the steepest slopes occur where valley rejuvenation has been most pronounced, and the oldest portions of the valley-walls seem to be significantly less steep than the more recently cut slopes.

3:3:4 Summary

In the preceding section it was pointed out that a large portion of the Whitemud basin valley lengths are synonymous with the channel system; this being particularly evident in the southern half of the

¹²The Student t test was not applied in this case because the assumption of homogeneous population variances was not satisfied.

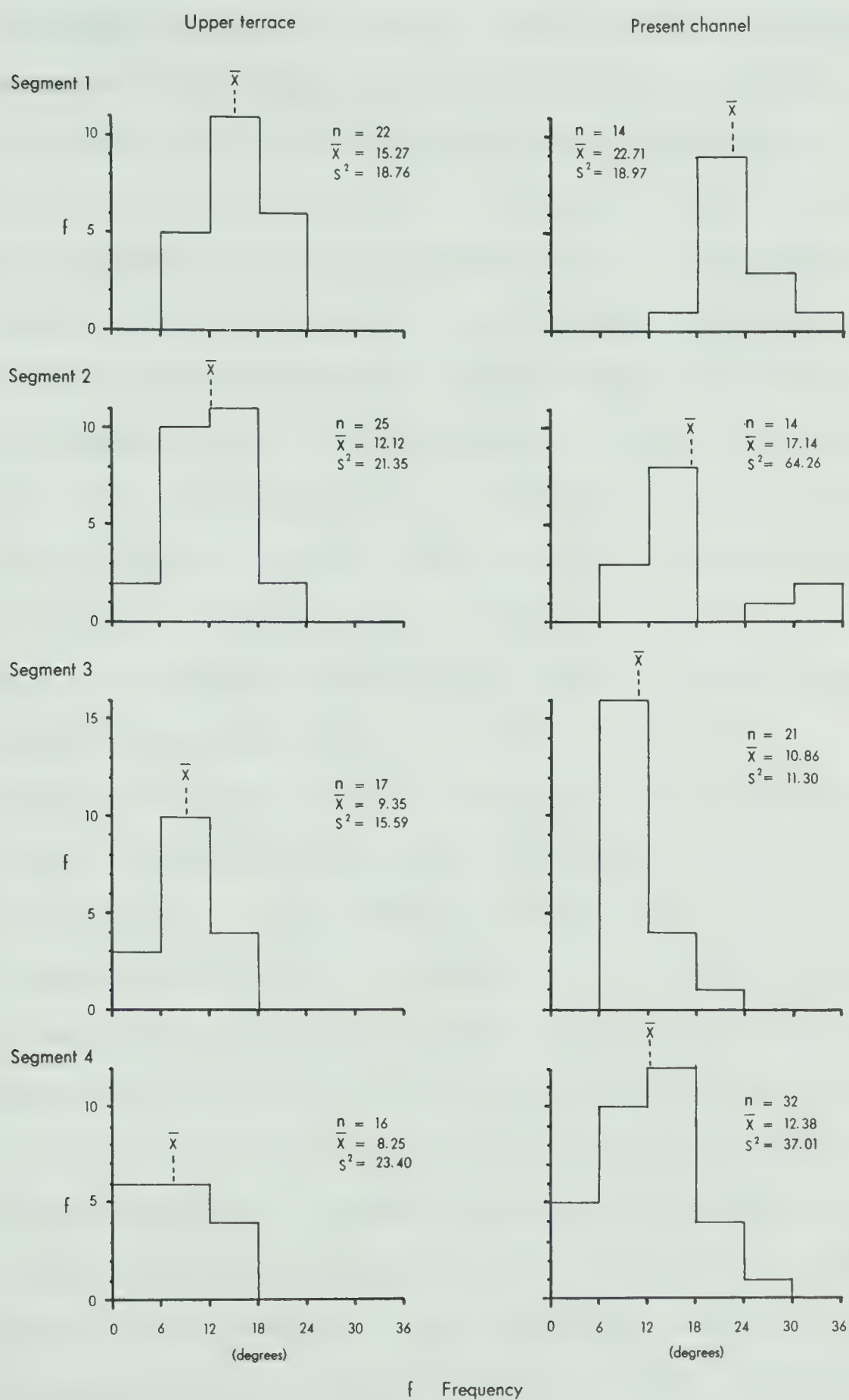


Figure 66 - Valley-wall Slopes

basin. Little detailed work was done on the channel/valley units in this area because land-use activities, in the form of stock-trampling and grazing, artificial damming, and the construction of roadside drains, have caused major modifications of natural form-elements.

Such is not the case, however, in that part of the valley between the Leduc/Calmar highway and the basin mouth. Consequently, this zone was a focus of attention. A well-defined sequence of three matched terraces was discernible and detailed mapping of surface elevations and description of alluvium exposures confirmed this view. It is possible that a major aggradation period took place prior to the upper terrace formation in that at least two unpaired remnants, consistent in terms of elevation, and of floodplain deposits, were encountered. It is feasible that subsequent phases of channel down-cutting removed the greater part of this old alluvial fill.

Analysis of terrace alluvial stratigraphies was based on simple descriptions of observed sections and on the mechanical analysis of fine-grained samples. In each terrace the lowest bed is usually composed of gravel which in turn is succeeded by fine-grained alluvium. However, the vertical relations of texture within the latter unit do not always conform to the general fluvial model, proposed by Visher (1965), based on observations in many investigations of Holocene and ancient fluvial deposits. The upper terrace is distinguished by a layer of very fine sediments immediately above the gravel and underlying slightly coarser material. This constitutes a reversal of the expected sequence outlined by Visher (1965) and others. The middle terrace has a sequence comparable to the upper unit but the anomalous reversal is not so clearly evident. On the other hand, in the lower

terrace, fine-grained alluvium is comparatively homogeneous throughout.

The analysis of valley-wall slopes also proved fruitful. The valley-walls consist of a series of sub-units, each differing in relative age. It was confirmed by analysis of data derived from a random sample of 100 slope profiles for each wall that the two slope populations are not significantly different. In other words valley asymmetry as noted, for example, by Harman (1968), is improbable in this case. A significant difference in average slope angles of the oldest and youngest sub-units (the latter being subjected at present to basal sapping by the Whitemud channel) is probably indicative of slope decline above the upper terrace (oldest sub-unit). Schumm (1967b) reviewed theories of slope evolution and concluded that such slope decline is favored where creep is the major erosion process and the accumulative zone at the slope-base is little affected by removal of material. Such conditions are favored in the Whitemud valley where a forest vegetation cover inhibits slumping and rainwash, and the relatively flat upper terrace remnant surfaces limit the removal of debris.

The emergence of mean average slope angles between 13 degrees and 15 degrees for many of the data sets may be coincidental but it is also possible that this range represents a form of equilibrium angle, produced where basal sapping is present or only recently terminated. In the lower part of the valley, where rejuvenative effects have been most pronounced, the steepest slopes occur. With increasing distance up the valley a decreasing average slope angle is evident.

3:4 Channel Characteristics

3:4:1 Introduction

Some mention has already been made of channel properties of the Whitemud basin, in particular the length and gross sinuosity parameters. A brief description of other channel attributes, where valley and channel are synonymous, has also been presented. However, specific site characteristics are also pertinent and this section is devoted to a description of channel reaches and sites selected from the main third order and fourth order channels in the basin. Attention was concentrated on relatively small segments of the channels but it is possible to make qualitative generalizations on the basis of such data.

At the outset it should be noted that while the major channels have sinuous courses, successive, well-developed, symmetrical meanders are rare. Where the channels are incised wholly into alluvium or Glacial Lake Edmonton sediments meander symmetry is encouraged. For much of the fourth order channel length, however, the channel banks are alternately alluvium and bedrock on opposite sides. While the bedrock is not particularly resistant to erosion it seems that differences in erodibility of the alluvial fill and bedrock are sufficient to inhibit meander symmetry over considerable distances. Such channels might be termed irregularly sinuous and are not suitable for the application of standard measures of meander plan geometry (see: Bates, 1939; Leopold and Maddock, 1953; Leopold and Wolman, 1957).

So that some estimation of channel dimensions could be arrived at, the field programme was orientated toward the following objectives:

- (1) the measurement of various cross-sections of the major channels to indicate variability in channel cross-section geometry and to establish a measure of cross-section central tendency
- (2) the description of riffle and pool bed-forms in selected reaches
- and (3) the measurement of grain-sizes typical of the riffle surfaces in the selected reaches.

Such data satisfy purposes of descriptive geomorphology and reflect, albeit generally, process-elements of the stream system.

3:4:2 Methodology

The sampling method for selection of cross-section sites was as follows: points were located at one-mile intervals along the fourth order channel to the first major bifurcation point. Thereafter the fourth order channel was sampled at two-mile intervals. The four major third order channels were divided at intervals of 3 miles. The primary reason for concentrating measurements in that length of the Whitemud Creek downstream of the first major bifurcation point was that this part of the drainage system represents the culmination of discharge from the entire basin network. Accordingly, it was considered that by taking a comparatively large number of measurements here a good estimate of average channel cross-section dimensions could be obtained.

No attempt was made to randomize the specific site selection in the field. Rather, sites with clearly-defined channel margins were chosen as close as possible to the chosen location. Thus, the upper limit of a slip-off slope, as denoted by vegetation, slope and

sediment differences, was taken as marking the cross-section horizontal plane. Where no slip-off slope existed the lowest cut-bank lip determined the cross-section upper boundary. This reference plane was levelled to a steel tape stretched across the channel, normal to the banks, either by measurements above water in pools, or if no water was present, with an Abney level. At sites where cross-sections were relatively wide depth measurements were taken at intervals of 2 feet along the tape. Narrow channels were measured at 1 foot intervals, and the channel width was read to the nearest 6 inches.

It is considered that this method of cross-section delimitation approximates the specific channel cross-sections at or near bankfull-discharge stages of streamflow. This is only an approximation, however, because in some places secondary, high-level channels are utilized during floods. A few yards from such sites only a single channel may exist. Such difficulties have often been encountered in studies of channel morphology and no simple answer to the problem has yet been attained (Neill, 1964). In this study the lowest discernible channel was measured at each site. Bankfull-discharge for such locations may not necessarily be the same for contiguous cross-sections. This problem is almost insurmountable unless detailed field measurements can be made simultaneously at various sites during different flow stages. With the above reservation the data are sufficient to indicate general cross-section dimensions.

The measurements of pool and riffle phenomena were originally based on a systematic selection of sites. To this end it was decided to measure five consecutive riffles and pools every 3 miles up the

main channel to the third order bifurcation point previously mentioned. Thereafter, sites were chosen every 6 miles along the remaining portion of the fourth order channel and along the four main third order channels. However, this rigid sampling plan had to be considerably altered for practical reasons. Cattle-trampling, beaver-dams and man-made dams severely disrupted natural bed-forms in all but the lower fourth order channel segment. Where large beaver-dams exist it is not possible to distinguish riffles, and cattle-trampling of dry channels completely alters the natural configuration. As a consequence only those sites ostensibly little-disturbed were incorporated. The following criteria were used to identify riffles for measurement:

- (1) Debris had to extend at least half-way across the channel.

Smaller accumulations were arbitrarily designated as incipient riffles and were ignored.

- (2) The transition between pool and riffle was subjectively demarcated on the basis of grain-size change and/or breaks in bed slope.

It is clear that exact replication by independent investigators is not obtainable for such work. Leopold, Wolman, and Miller (1964) commented on the subjective decisions necessary in distinguishing riffle margins but they did not consider this a serious deficiency.

A modification of the Wolman (1954) sampling method was employed in the measurement of gravel-size on the riffle surfaces. Detailed size analysis usually entails the measurement of three axes of individual stones but for this exploratory study only the long-axis was measured. A steel tape was laid on the riffle surface along three

equally-spaced lines parallel to the riffle cross-channel axis. Stones falling under each foot division were measured. Where riffles were formed of fine-grained sediments two spot samples were taken from each riffle for subsequent mechanical analysis.

3:4:3 Whitemud Data

The fourth order channel longitudinal profile, from the basin mouth to the Leduc/Calmar highway is illustrated in Figure 67. Depicted as well are cross-sections measured along this length and average channel slopes for major facets of the longitudinal profile. Form ratios, defined as the average cross-section depth divided by the channel width (Morisawa, 1968), calculated for each of the cross-sections were grouped into frequency histograms (Figure 68). For this purpose the cross-section data were placed in three groups:

- (1) those of the fourth order channel segment downstream of the first third order bifurcation point
- (2) those of the remaining portion of the fourth order channel
- and (3) those of the third order channels.

While it is not permissible to calculate sample variances and to apply the difference of means test when using systematic samples such as these (Spence, 1969), it is reasonable to qualitatively compare sample means if the frequency distributions are not markedly skewed. The form ratios shown in Figure 68 indicate that the lower fourth order channel segment has relatively wide and shallow cross-sections. The remaining portion of the fourth order channel has an average form ratio comparable to that calculated for the third order channels. Here the channel depths are greater in proportion to the channel widths.

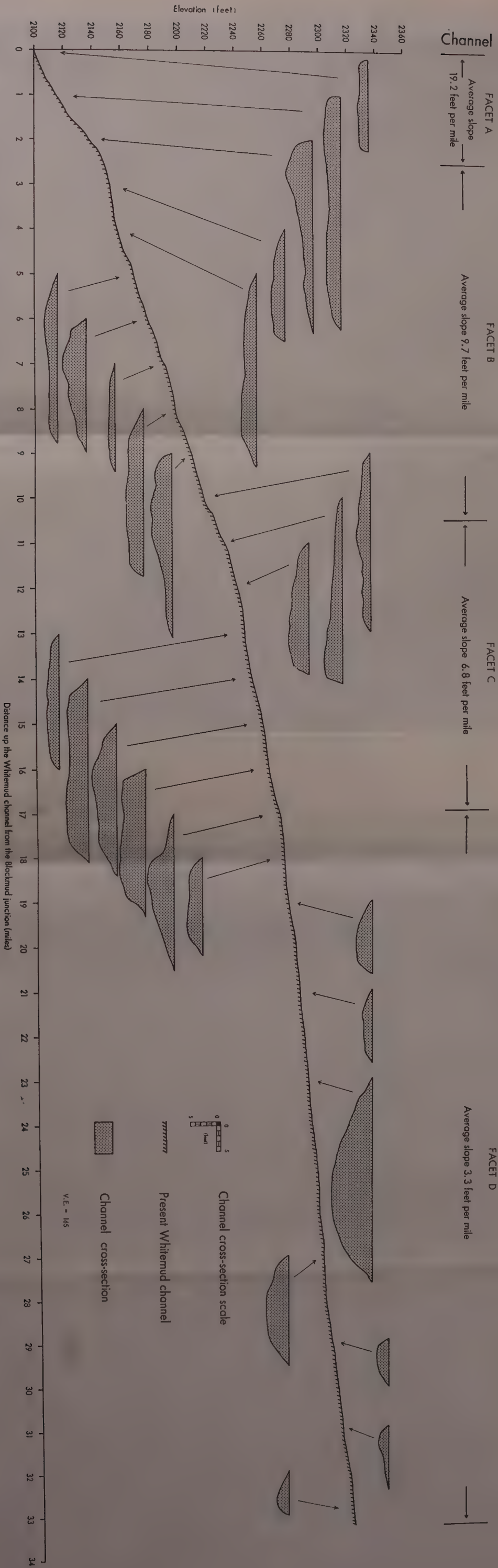
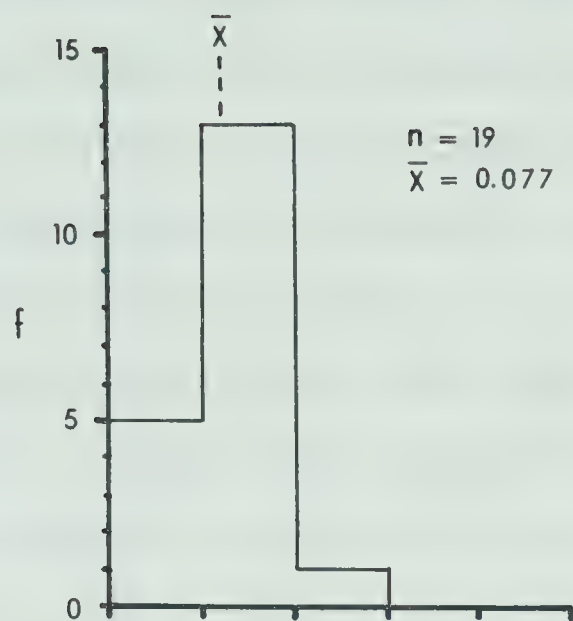
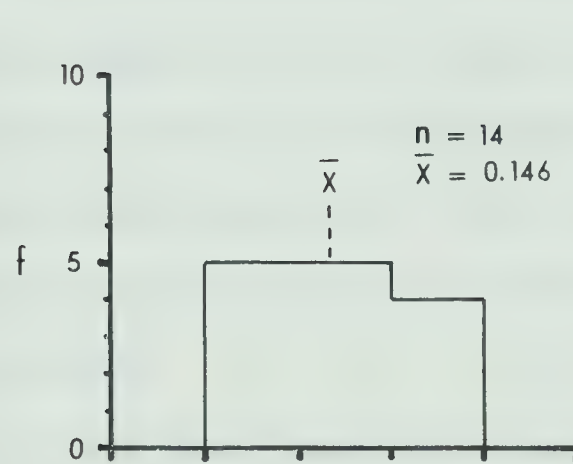


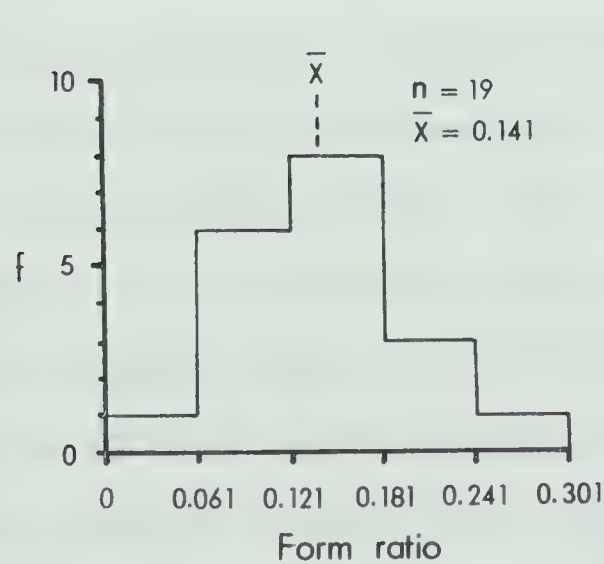
Figure 67 - Partial Longitudinal Profile and Selected Channel Cross-sections of Whitemud Creek



Primary segment of fourth order channel



Secondary segment of fourth order channel



Third order channels

f Frequency

Figure 68 - Form Ratios of Channel Cross-sections

Many of the channels revert to wide, shallow forms again in their head-water reaches, especially where grassed-waterway channel forms predominate. Cross-sections were not measured at these sites because of the extreme subjectivity involved in delimiting the channel margins.

It is interesting to note that the relatively wide, shallow channel cross-sections of the lower fourth order channel segment occur where the bed is comparatively resistant to erosion. The channel here is cut into bedrock which is mantled quite uniformly with a gravel veneer. The banks of fine-grained alluvium and bentonitic sandstone and shale probably are more easily eroded and the channel, therefore, widens at the expense of depth. Conversely, the remaining cross-sections are for the most part cut into fine-grained alluvium and fine-grained Glacial Lake Edmonton sediments. Here the bed-resistance is perhaps not so marked and the channels are scoured more easily, producing higher form ratios.

In passing, two comments might be made regarding the form of the Whitemud Creek longitudinal profile (Figure 67). First, the profile has little tendency to approach the classic concave-upward "graded" form but, rather, has the over-steepened lower reaches typical of rejuvenated systems (Chorley, 1958b). Structural elements also appear to have played a part in determining this profile. For example, the break in slope between Facets A and B (Figure 67) coincides with a very resistant clay-ironstone stratum, up to 2 feet thick, which is visible in the channel bed and adjacent terrace exposures at that locality. Similarly, the break in slope between Facets B and C occurs where bedrock and glacial till cease to be an influence on the channel form

(Figures 49 and 67). Thus, periodic rejuvenation and structural control largely explain this longitudinal profile.

The location of each site selected for riffle and pool measurement and the distribution and dimensions of pools and riffles in five of these channel reaches are shown in Figures 69 to 72. There appears to be no systematic spacing of the riffles as observed by Leopold, Wolman and Miller (1964). They concluded that riffles occur successively at distances five to seven times the channel width but this is only partially true of the Whitemud examples. Riffles at site 1 are separated on the average by a distance four times the mean channel width; at site 2 the spacing is twelve times the width. Sites 3, 4 and 5 have spacings of ten, six, and five times respectively. However, it is unlikely that the Whitemud riffle sequence is highly abnormal because incipient riffles were ignored, as were artificial accumulations of organic debris which may serve, at least temporarily, as minor riffles during the streamflow period. Leopold, Wolman and Miller (1964) noted that pools coincide with meander bends because the intensity of erosion is greatest at those points. The local riffles conform well to this observation with the inflection points of meander paths being represented by riffle concentrations (Figures 70 to 72).

Analysis of riffle grain-size data (Figure 73) showed that in the lower 14 miles of the Whitemud channel riffles are composed almost exclusively of gravel. Thereafter riffles are generally formed of fine-grained sediments. Pools in the lower 14 miles of channel also have a gravel veneer with only minor amounts of fine material. The average length of gravel long-axes at each site varies slightly from

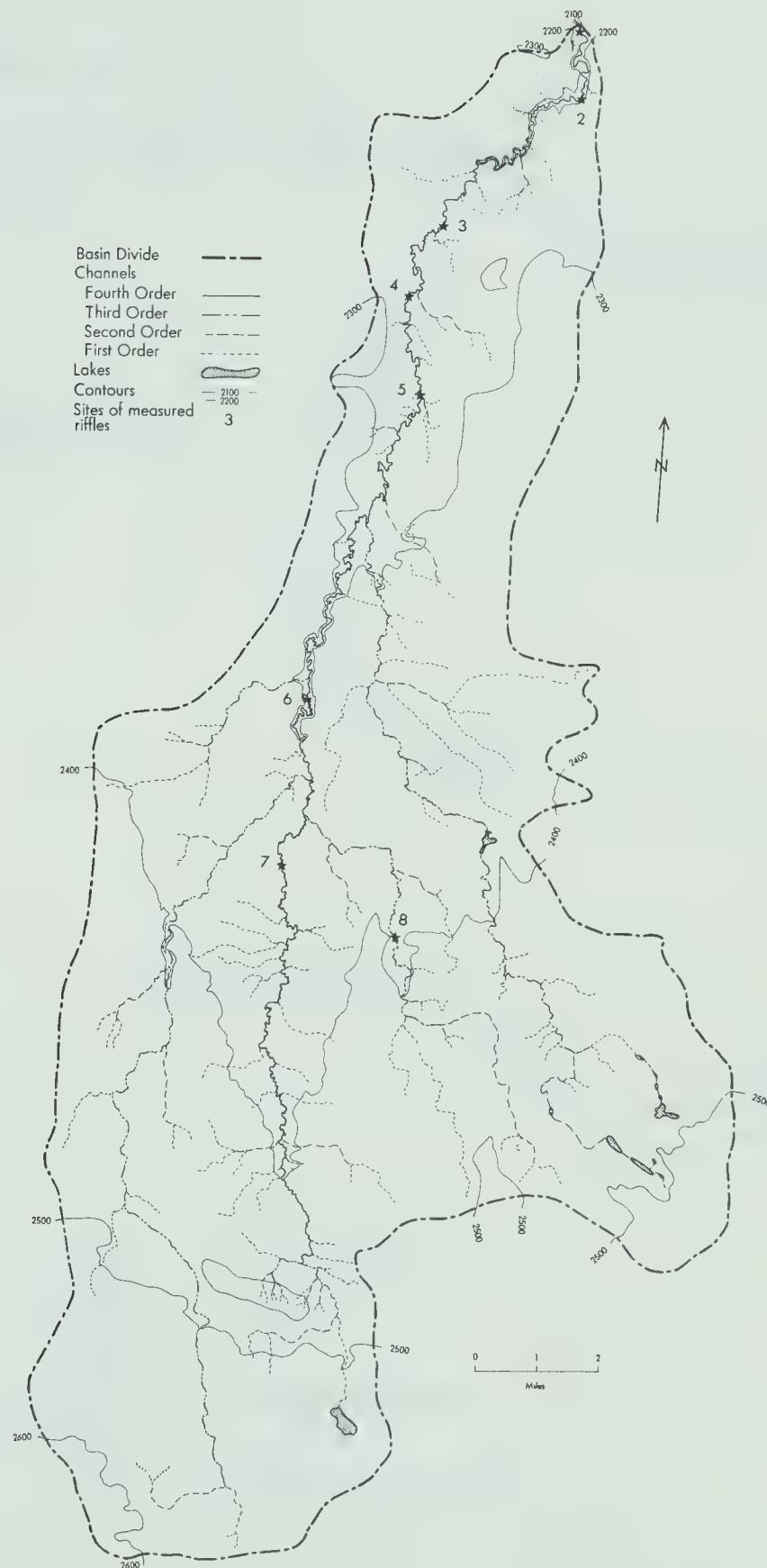


Figure 69 - Locations of Measured Riffle Sites

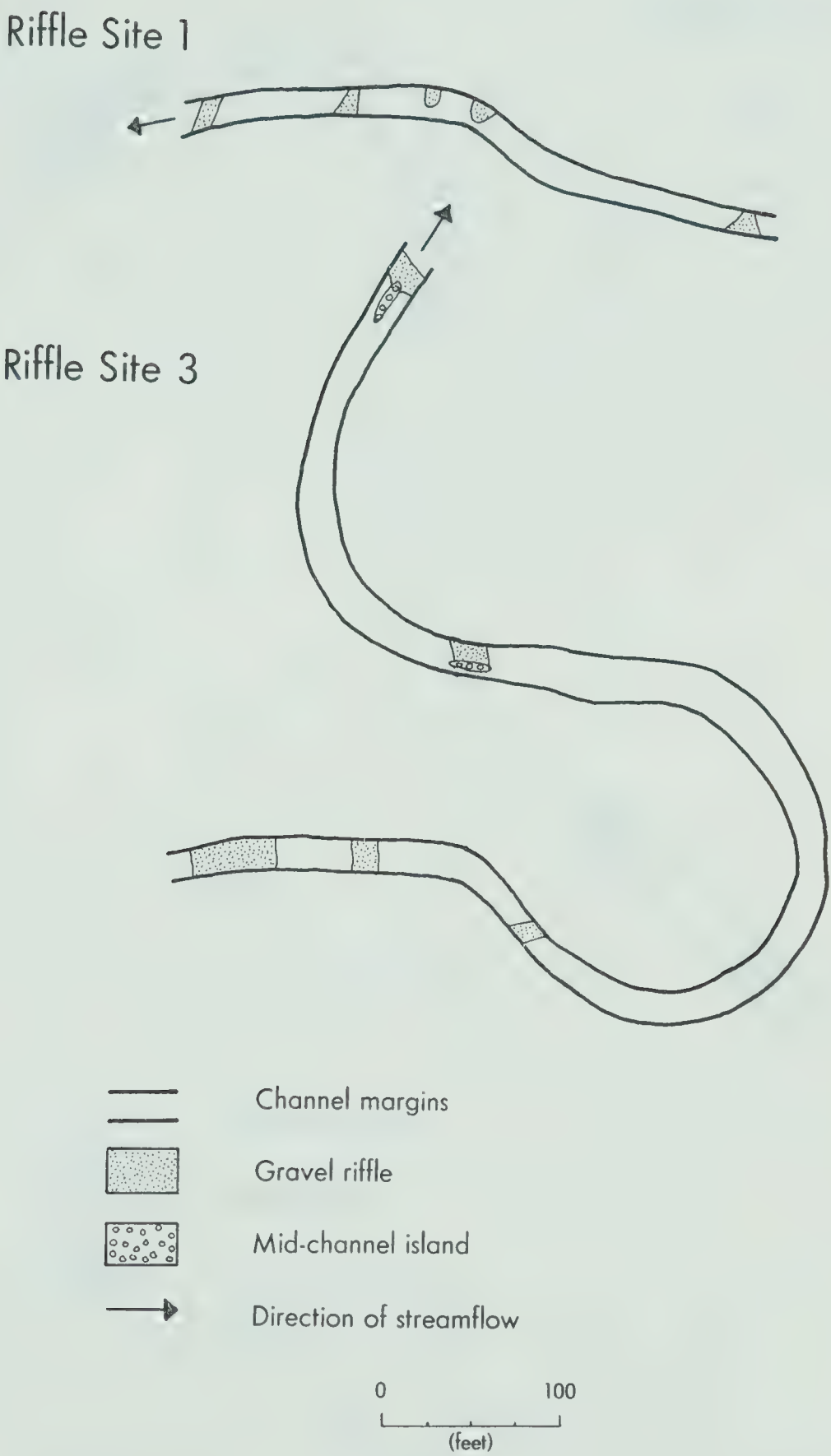
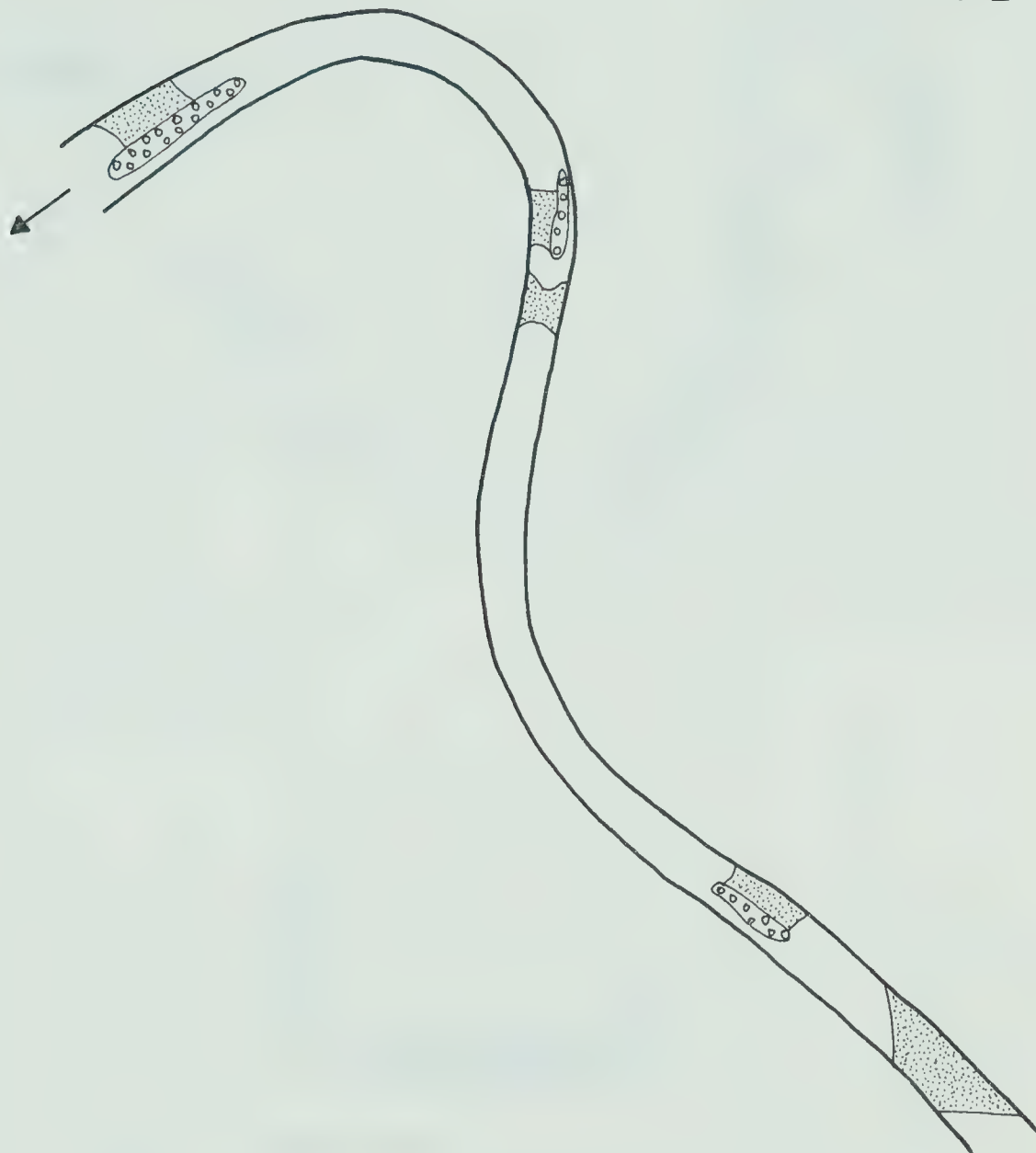


Figure 70 - Riffle Sites 1 and 3

Riffle Site 2

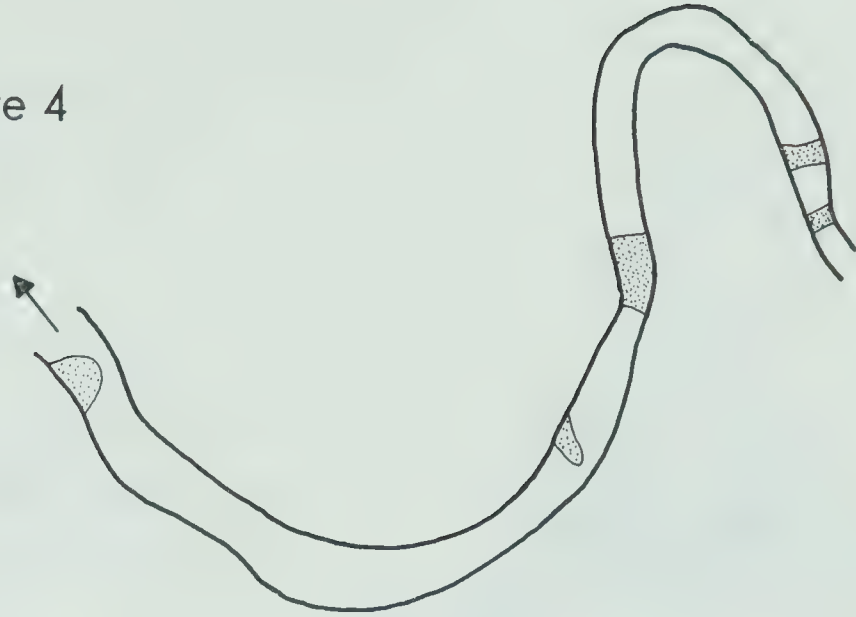


-  Channel margins
-  Gravel riffle
-  Mid-channel island
-  Direction of streamflow

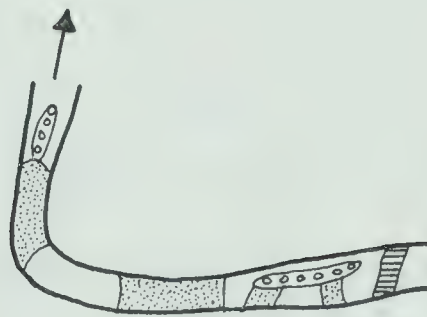
0 100
 (feet)

Figure 71 - Riffle Site 2

Riffle Site 4



Riffle Site 5



-  Channel margins
-  Gravel riffle
-  Mid-channel island
-  Fine-grained riffle
-  Direction of streamflow

0 100
 (feet)

Figure 72 - Riffle Sites 4 and 5

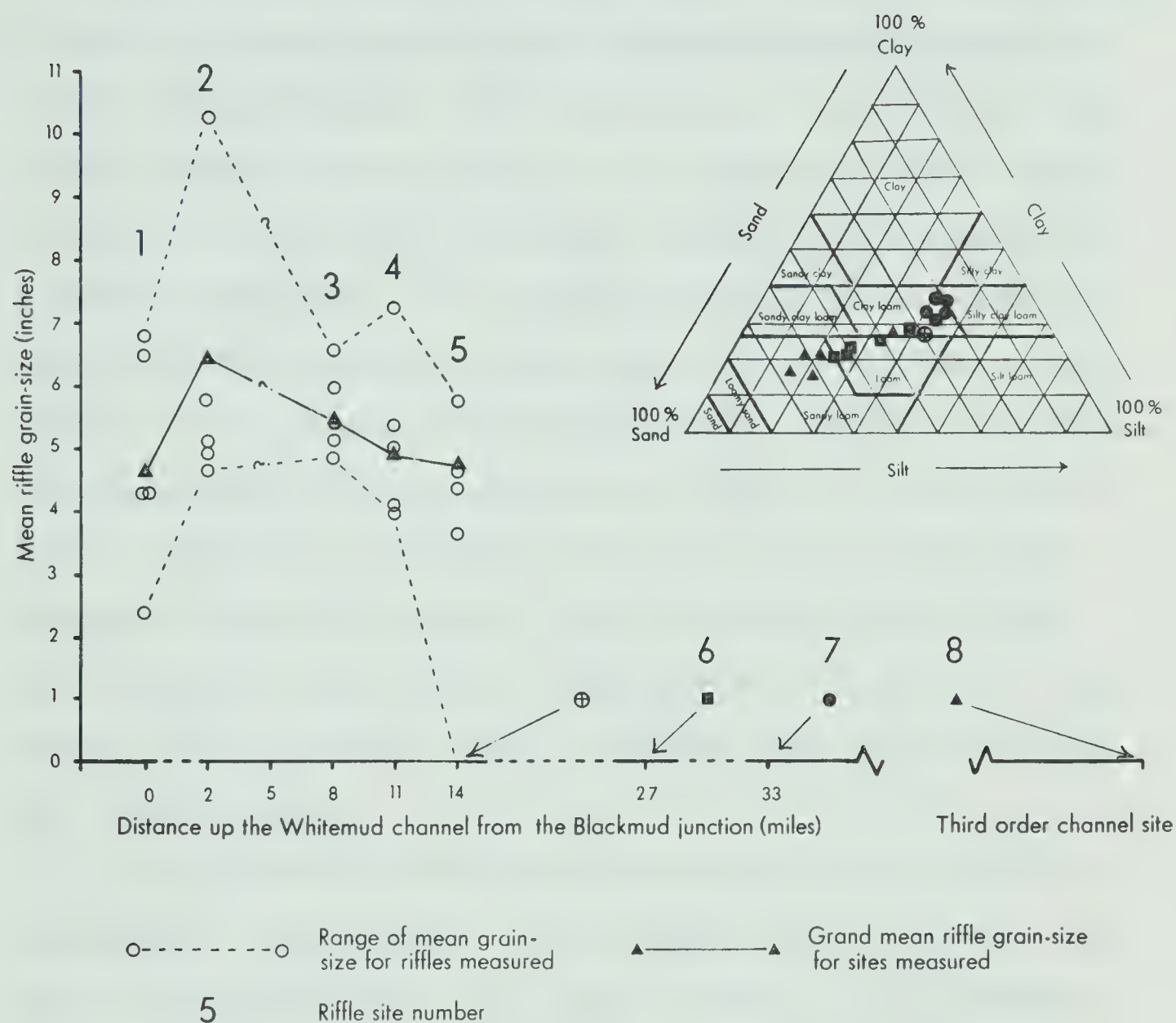


Figure 73 - Riffle Grain-size Data

about 4.5 inches at the basin mouth through 6.5 inches two miles upstream, to around 5.0 to 5.5 inches at sites 3, 4, and 5.¹³ The riffle farthest upstream at site 5 contrasts with others in the vicinity in that it is formed exclusively of fine-grained debris washed out of a small tributary channel. This abrupt change in dominant grain-size cannot, therefore, be attributed to some threshold mechanism in the streamflow. Nevertheless, this general area appears to coincide with a change in deposition. Pools upstream from this locality contain a few stones which apparently became trapped during high flows. Alternating with the pools are fine-grained riffles. Extreme difficulties were encountered in finding undisturbed riffles in this part of the basin. Consequently, only three sets of fine-grained riffles were sampled for mechanical analysis. While contiguous riffles in each set show certain consistencies in the amount of silt, sand, and clay (Figure 73), no systematic variation upstream could be discerned from such a small sample.

One fundamental generalization regarding the size distribution of bed material may be made. The existence of gravel as the dominant fraction in certain parts of the channel system is clearly dependent on the availability, at close quarters, of such material during channel erosion. The three main sources are: indurated bedrock strata, glacial till beds, and the gravel at the base of alluvial terrace de-

¹³Only the four gravel riffles were used to calculate the grand mean for site 5.

posits. It seems that the coarsest gravel, once entrained in the channel moves only short distances downstream. For example, very large slabs of relatively resistant sandstone form a number of large riffles 1.5 miles upstream from the basin mouth. These slabs are contributed by rockfall and slumping from a massive sandstone stratum exposed about 30 feet above the channel. Fifty yards downstream there is little trace of this material in the channel bed, except for small fragments a few inches long. Therefore, although sorting and redistribution of the smaller gravel must take place during peak flows a large portion of the bed material may move only rarely and for short distances. The importance of gravel sources in close proximity to the channel may be illustrated further; the transition from gravel to fine-grained riffles occurs precisely in the area where bedrock, till, and alluvial gravel are not yet exposed in the channel (Figure 49). The Glacial Lake Edmonton sediments are not a good gravel source and it is, therefore, impossible for a large mass of coarse material to accumulate in those parts of the channels not incised through the glacio-lacustrine deposits.

3:4:4 Summary

Although it was possible to sample only small parts of the main Whitemud channel for detailed study, and the lower order channels received even scantier attention, four main conclusions emerged.

First, the channel long profile, from the basin mouth to the Leduc/Calmar highway seems to have been determined largely by periodic rejuvenation, which has over-steepened the lower reaches. A secondary element of structural control is evident at two of the more pronounced

breaks in slope. Of note in this regard is the fact that the generalized terrace surfaces tend to follow closely the present channel long profile. This prompts the suggestion that the two main structural anomalies (a thick clay-ironstone stratum, and the transition from till to glacio-lacustrine sediments) have been effective throughout the greater part of the valley-cutting period.

Second, the analysis of channel cross-sections revealed a three-fold division of forms. Form ratios are relatively low in the main segment of the Whitemud channel up to the first third order bifurcation point. Third order channels and the remaining fourth order segment tend to have slightly higher values. It is thought that this difference is a function of variable bed resistance, the gravel-covered bedrock of the fourth order segment being more resistant than the essentially non-graveliferous reaches upstream. Further upstream, near the headward limits of most channels, very wide, shallow forms exist. It appears that the vegetative cover has been sufficient to prevent rapid channel-cutting in these areas, even in the years with relatively large runoff. Exceptions, of course, are those channel-reaches manifestly disturbed by land-use activities.

Third, the mapping of riffles and pools in selected reaches indicated that the spacing of riffles is not as regular as that noted by other investigators. However, this may be more apparent than real in that an arbitrary size limit determined the inclusion or exclusion of debris accumulations.

Fourth, variation in grain-sizes on the riffle surface with increasing distance upstream is closely related to the immediate geologic

environment. Gravel from a number of sources remains in the lower main channel reaches and only the smaller fragments appear to be re-distributed far downstream. From a point about 14 miles upstream, where gravel becomes scarce, the bed is composed mainly of fine-grained debris. Very minor amounts of sand, silt and clay are deposited in pools near the basin mouth. Such sediments entrained during the spring runoff period must be flushed almost entirely out of the system. However, relative to streams in other climatic regions, streams in this part of the world carry very small loads per unit area (Morisawa, 1968). Therefore, while the present channel is probably in a degradation phase such down-cutting must be extremely slow because;

(a) peak flow occurs for a very short period each year

and (b) the gravel veneer appears to be comparatively stable and must protect the weakly-resistant underlying bedrock from rapid erosion.

3:5 Summary of Whitemud Interfluvial, Valley, and Channel Morphologic Character

This chapter has centred on the analysis of data collected mainly in the field. Underlying this analysis was the primary objective of explanatory description of interfluvial, valley, and channel landforms.

The interfluvial areas were analyzed in terms of maximum slope angles, lengths, and orientations at selected point locations. Slope angles are generally very low and there is little difference between mean maximum angles in either lacustrine or non-lacustrine surficial geology zones. However, the range and dispersion around the mean angles is much greater in areas of non-lacustrine material. Thus, the

steeper slopes tend to occur near the basin margins and in the southern half of the basin where lacustrine sediments, if present, are thin. The preferred orientation of the measured slopes is difficult to explain. Localized surface depressions and undulations tend to counter the broader, gross slopes leading from the divide, tangent to the basin long-axis. This is illustrated by the relative absence of specific site slope orientations in the north-west sector.

Terraces and valley-wall slopes in the main Whitemud valley were studied in some detail. At least three paired aggradation surfaces were distinguished, with the possibility of a further higher surface, now largely destroyed. The associated stratigraphic sequences conform loosely to Holocene deposits investigated elsewhere but at least one major anomaly was identified. This is the superposition of clay and silt-rich sediments immediately above the basal gravel unit of two of the terraces. Such an occurrence is at variance with the general fluvial model presented by Visher (1965). Structural characteristics were briefly described but differentiation of the deposits was based mainly on the analysis of texture. It appears that the middle and upper terraces are comparable in terms of texture but have varied structural elements. The lower terrace differs from both the middle and upper units in texture and structure.

Average slopes are apparently similar for both the east and west walls of the valley. As a consequence of repeated rejuvenative phases the steepest slopes occur close to the basin mouth and become progressively less steep up-valley. There is also some evidence to indicate that slope erosion processes have partially transformed the

oldest portions of the valley-walls. It is possible that these have declined in the order of 3 degrees since they were cut by lateral wandering of the related stream.

The irregular sinuosity of the main Whitemud channel seems to be determined mainly by differential bank resistance to stream erosion, that is, bedrock and alluvium are eroded at unequal rates. Meander symmetry is best-developed further upstream where the channel is incised in more homogeneous material. Measurements of channel cross-sections led to the suggestion that subtle changes in cross-section form are related to the bed resistance. This resistance, which is at a maximum in the lower reaches of the Whitemud channel, is increased by a uniform veneer of gravel overlying the bedrock into which the channel is cut. Further upstream, where bedrock and gravel cease to form the channel bed, the cross-sections deepen relative to width. In the headwater environs of most channels very wide and shallow cross-sections are probably related to low discharge. The relatively small and intermittent amounts of water carried along these reaches are unable to severely disrupt the vegetation and promote channel incision. Exceptions occur, however, where land-use activities upset the natural conditions.

The partial longitudinal profile constructed for the Whitemud channel illustrates the elements of rejuvenation and structural control. Bedforms in various reaches were measured in terms of plan dimensions and grain-size distributions. The distribution of grain-sizes proved especially interesting. Gravel of riffle surfaces in the lower reaches of the main channel tends to have average long-axis lengths in the order of 4 to 7 inches. Much larger material occurs in places where

lithologic conditions permit. In general, it seems that the relatively large debris moves only over very short distances downstream. Finer gravel probably is redistributed over considerable distances during spring periods of peak stream flow. The relatively small variability in average gravel long-axis length over the lower 14 miles of the channel bears witness to this. Upstream of the 14-mile site riffles are usually formed of fine-grained sediments. The little gravel present here remains trapped in the pool bottoms. There is a clear relationship between the availability of gravel and the formation of gravel riffles and a channel gravel veneer which occurs even in pools. Where bedrock, till, and alluvial gravels are not exposed close to the channel, fine-grained sediments mantle the channel bed. However, further downstream, where gravel is available and maximum discharges take place fine sediments are almost entirely absent from the channel deposits.

CHAPTER FOUR

THE WESTERN PLAINS POSTGLACIAL ENVIRONMENT AND GEOMORPHIC HISTORY OF THE WHITEMUD BASIN

4:1 Introduction

A tremendous volume of literature on Pleistocene and Holocene geology, climate, botany, and archaeology has amassed in the last half-century. While controversies abound within each of these disciplines the main elements of climatic change during Holocene time seem firmly established. Early work by Brooks (1951) revealed parallel rates of ice sheet recessions at the end of the Wisconsin Stage in eastern North America and north-west Europe. He noted that peat bogs in both areas exhibit comparable evidence of alternating wet and dry periods and thus concluded that climatic changes were essentially synchronous. Recent work on peat bogs in Keewatin and northern Manitoba (Nichols, 1967) demonstrates the synchronicity of climatic change in north-west Europe and central Canada over the last 6000 years.

Antevs (1955) established a postglacial geochronology of the southwestern United States. His threefold division of postglacial (Neothermal) time, based on subdivisions of the July temperature curve, was as follows (adapted from Leopold, Wolman and Miller, 1964, p. 406):

- (1) Anathermal: approximately 10,000 to 7,500 years B.P.
- (2) Altithermal: approximately 7,500 to 4,000 years B.P.

and (3) Medithermal: approximately 4,000 years B.P. to the present. While this may serve as a framework for estimates of relative time Bryan and Gruhn (1964) cautioned against the indiscriminant use of it as an absolute dating scheme. Thus, they stated (Bryan and Gruhn, 1964, p. 309):

There is no doubt about the reality of the Neothermal (Anathermal-Altithermal-Medithermal) temperature sequence. The Neothermal temperature curve, with three sequent phases, from relatively cool to relatively warmer conditions like the present, has been demonstrated at a number of localities throughout the Northern Hemisphere and is now almost universally recognized (Cooper, 1958).However, as Cooper (1958) has rightly pointed out,...subdivisions of the Neothermal period are merely "phases" of the temperature curve and not time periods; and radio-carbon evidence...shows that absolute dates for these phases cannot be projected from one part of the world to another.

In a paleometeorological study Bryson and Wendland (1967, p. 287) noted that "one of the striking features of pollen diagrams from central North America is the abrupt transition from Late Glacial to post-Glacial pollen assemblages." They considered that a rapid change from boreal forest to grassland began about 10,500 years B.P. in southwestern Manitoba, immediately after the Valders Substage maximum. Following this there was a period of rapid glacier wasting, with the exception of a localized readvance near Cochrane, Ontario, approximately 8,500 years B.P. (Hughes, 1965). A continued amelioration of temperature led into the Altithermal period. Reviewing the radiocarbon dating (C^{14}) evidence for this warm, relatively dry period Bryan and Gruhn (1964) noted that the Altithermal interval probably lasted from nearly 8,000 to 3,500 years B.P. in southern British Columbia. Further, they mentioned

the work of Zumberge and Potzger (1955) which placed the time of maximum aridity in the Lake Michigan basin at approximately 4,000 years B.P. It is possible that the period of greatest aridity occurred much earlier in the northwestern United States. On the basis of the stratigraphic position of C^{14} dated material in water-laid sand and aeolian silt, Gruhn (1961) estimated that maximum aridity was attained about 6,500 years B.P. in south-central Idaho. Insofar as central North America is concerned changes in climate over the last 5,000 years may be considered as relatively minor deviations from present conditions (Bryson and Wendland, 1967).

4:2 The Western Plains Postglacial Environment

A summary of some of the work which has been done on the postglacial climate, vegetation, and geomorphology of Manitoba, Saskatchewan and Alberta follows. The emphasis has been placed on studies most relevant to the formation of postglacial alluvial terraces in Alberta.

Recently, Bryson and Wendland (1968) reconstructed the Laurentide ice margin positions at various times during the postglacial period. This involved an interpolation of isochrones from C^{14} dates considered indicative of postglacial deposits in central and northern North America. They were hampered by a small number of deglaciation absolute dates for Alberta and the isochrones for this area are tentative. Their interpretation was that the Laurentide ice margin lay immediately north or east of the present study area about 10,500 to 10,000 years B.P. (Bryson and Wendland, 1968, ff. p. 14). Other evidence, to be discussed later, points to ice marginal positions being much further north

The first part of the paper discusses the importance of the
theoretical framework in the study of the
relationship between the variables. The second part
presents the empirical results of the study. The third part
discusses the implications of the findings for the
theory and practice of the field. The fourth part
concludes the paper and suggests directions for
future research.

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or east at that time. Estimates of the ice sheet profile led them to hypothesize that an ice-free corridor between the Arctic and the plains did not come into existence until 9,000 to 8,000 years B.P. They considered that if the structure of Arctic air then was similar to the present, the influx of cold Arctic air after the formation of the corridor might have lowered the winter temperatures of the plains by as much as 20 degrees Centigrade.

Hansen (1949) made certain inferences about the postglacial vegetation sequence in central Alberta from his analysis of pollen profiles for a number of peat bogs. He noted that in most cases pollen of pine and spruce predominated, with secondary elements of balsam fir, aspen poplar, grasses, composites, and chenopods. The grasses, composites, chenopods, and pines attained their maximum representation in the lower middle portions of the pollen profiles. Taylor (1969) thought that the grassland complex developed to a peak between 5,700 to 5,000 years B.P. in the Peace Hills area, near Wetaskiwin. Faunal remains at the Peace Hills archaeological site indicated a change toward the present Parkland vegetation about 4,000 years B.P. (Taylor, 1969, p. 91).

Evidence of vegetation change in southern Manitoba largely corroborates the general sequence established by Hansen (1949). Thus, Ritchie (1967, p. 225) stated:

...the Glenboro [Manitoba] site suggests a succession of floral assemblages starting with a mixed assemblage dominated by spruce with temperate and boreal associates... next an assemblage suggesting a grassland, then a zone interpreted in terms of deciduous forest, and finally a change in the late Holocene induced by settlement.

Radiocarbon dates from materials buried in postglacial alluvium in the three prairie provinces are presented in Table XVI. The sparse number

TABLE XVI - SOME RADIOCARBON DATES FROM POSTGLACIAL ALLUVIUM IN MANITOBA, SASKATCHEWAN AND ALBERTA

Location and Sample Code	Dated Material and Site	Date (years B.P.)	Source
Cochrane, Alberta GSC - 613 GSC - 612	Bones, from a Bow River middle post-glacial terrace. (as above).	11,370 $\pm$ 200 10,760 $\pm$ 160	Lowdon, Fyles & Blake (1967)
Taber, Provincial Park, Alberta S-68 GSC - 3	Wood, from early postglacial alluvium of Old Man River. (as above).	11,000 $\pm$ 250 10,500 $\pm$ 200	McCallum & Dyck (1960) Dyck & Fyles (1962)
Holland, Manitoba GSC - 492	Freshwater clam shells, from terrace 50 feet above Assiniboine River.	10,670 $\pm$ 160	Lowdon, Fyles and Blake (1967)
Duffield, Alberta GSC - 767 S - 106 S - 107 S - 140	Wood, from top of river gravel beneath alluvial clay and marl. Wood, at base of marl deposited on lowest N. Saskatchewan River terrace. Wood, 10 feet above base of marl. Marl, 11 feet above base of marl.	8,320 $\pm$ 140 8,150 $\pm$ 100 7,350 $\pm$ 100 10,600 $\pm$ 300*	Lowdon & Blake (1968) McCallum & Wittenberg (1962)
Edmonton, Alberta I - 2612 I - 2778 (not yet pub.)	Charcoal, 9 feet below surface of lowest N. Saskatchewan River terrace. Charcoal, 7 feet below surface of the above terrace. The A horizon of a humic gleysolic palaeosol near top of lowest N. Saskatchewan River terrace.	6,630 $\pm$ 200 6,290 $\pm$ 250 5,000	Westgate (1969) IN Westgate, Smith & Nichols (1969) as a personal comm. from Dr. S. Pawluk, Dept. of Soil Sci., U. of Alberta
Russell, Manitoba GSC - 280	Wood, from an Assiniboine River terrace.	6,320 $\pm$ 140	Dyck, Fyles and Blake (1965)
Oxbow, Saskatchewan S - 44	Charcoal, 9.5 feet below surface of Souris River lowest terrace.	5,200 $\pm$ 130	McCallum and Dyck (1960)
High River, Alberta L - 391H	Wood, from 1 foot above base of 15 feet of floodplain silt overlying gravel in abandoned channel of Little Bow River.	4,200 $\pm$ 150	Olson and Broecker (1959)
Winnipeg, Manitoba GSC - 216 GSC - 215	Wood and freshwater shells, from bank of Red River. (as above)	3,660 $\pm$ 130 3,650 $\pm$ 140	Dyck, Fyles and Blake (1965)
Steveville Dino-saur Park, Alberta GSC - 443	Wood, 12 feet below surface of Little Sandhills Creek floodplain.	360 $\pm$ 130	Dyck, Lowdon, Fyles & Blake (1966)

* This date was considered to have been distorted by alteration of the C^{14} content of the marl (McCallum and Wittenberg, 1962).

of available dates has thus far prevented the establishment of a detailed terrace chronology as has been possible, for example, in the south-west of the United States (see: Bryan, 1925; Leopold and Miller, 1954; Miller and Wendorf, 1958). Nevertheless, the dates are suggestive of the major periods of aggradation.

Certain implications of some of these dates were noted by the investigators who collected the material. Thus, Klassan (in Dyck, Fyles, and Blake, 1965) stated that the Winnipeg dates fall within a Red River aggradation phase known to have begun about 6,750 years B.P. (Elson, 1962), and that the Russell date refers to a late phase in the filling of the Assiniboine River valley. In relation to the Duffield dates it was observed (McCallum and Wittenberg, 1962) that the North Saskatchewan River must have down-cut rapidly after deglaciation to almost its present level. A recent confirmation (GSC-767, Table XVI) of the earlier Duffield terrace dating (S-106, S-107, S-140, Table XVI) was noted by Hills (in Lowdon and Blake, 1968). An early postglacial period of rapid down-cutting was also envisaged for the Souris River, near Oxbow, Saskatchewan (Nero and McCorquodale, 1958). Finally, it should be pointed out that the Steeville date was derived from a badlands floodplain where rates of aggradation are greater than usual.

Three detailed studies of postglacial terraces in Alberta merit close attention. Stalker (1968) described a number of terraces at Cochrane, in the Bow River valley. He distinguished five postglacial terrace levels, the lowest one consisting of the present floodplain and slip-off slopes. It was considered that a large mass of valley fill was deposited here approximately 12,000 to 10,000 years ago. Following

this the Bow River "carved the five lower terraces from [the] fill and... underlying bedrock" (Stalker, 1968, p. 1455). A most important point is that he thought the alluvium of at least the upper four terraces was deposited during one period of aggradation. Thus, Stalker (1968, p. 1461) stated: "It appears that this deposition was fairly continuous, once under way, and was interrupted neither by any major hiatus nor by any extended period of erosion."

McPherson (1968), in a study of the lower Red Deer valley, described a number of low unpaired terraces similar in composition and structure to the sediments of the modern floodplain. He also noted unpaired sand and gravel terrace remnants at generally higher elevations in the valley. The lower terraces, with edges ranging in height from 4 to 15 feet above the floodplain, were considered to have been formed in the past several thousand years. The upper units, located about 100 feet above the Red Deer River, were tentatively dated as being older than 7,000 years by the presence of an elephant skull in the alluvium, near Empress (McPherson, 1968, pp. 231 and 234).

Recent work (Westgate, 1969) in the North Saskatchewan River valley at Edmonton is particularly pertinent because the Whitemud basin is located in the near vicinity. Four postglacial terraces were recognized and although the lower two were studied in detail it was noted that little is yet known about the upper units. He correlated the second terrace above the present channel with the Cochrane valley fill of 12,000 to 10,000 years B.P. (Stalker, 1968) on the basis of similar faunal remains. Radiocarbon dates derived from material in the lowest terrace (S-106, I-2612, and I-2778, Table XVI) placed this aggradation phase between

8,500 and 5,000 years B.P. A further confirmation of the age of this terrace was provided by the discovery of a Mazama ash marker-horizon, dated elsewhere at 6,600 years B.P., at the Beverly site (Westgate, Smith and Nichols, 1969). Recent degradation to the present channel level was assigned, therefore, to the last 5,000 years (Westgate, 1969). The estimates of depositional and erosional phases in Alberta during postglacial time, as presented by Stalker (1968), McPherson (1968) and Westgate (1969) are shown in Table XVII.

4:3 The Whitemud Basin Geomorphic History

It has not been firmly established when deglaciation of the study area took place. Bryson and Wendland (1968) indicated that ice may have been present here as late as 10,000 years ago. If, however, Westgate's (1969) suggestion that a major period of North Saskatchewan River valley degradation occurred between 12,500 and 11,500 years ago is correct then the ice margin must have lain well to the north or east at that time.

Bayrock and Hughes (1962) did not attempt to date the development and history of Glacial Lake Edmonton. Although much remains unknown about this phase they thought the lake probably took only a short time to reach its maximum development. They also believed that the inundated Edmonton area was almost completely drained, over a period of about 40 years, by the Gwynne Channel.

The early postglacial history of the North Saskatchewan River in this area was marked by the deposition of sandy alluvium over a large portion of the Glacial Lake Edmonton sediments. Following this "one of the anastomosing channels was enlarged and incised to become the present

TABLE XVII - THREE ALBERTA POSTGLACIAL ALLUVIAL TERRACE SEQUENCES

Postglacial time (years B.P.)	North Saskatchewan River valley (Westgate, 1969)*	Bow River valley (Stalker, 1968)*	Lower Red Deer River valley (McPherson, 1968)*
	?	?	?
13,000	Major <u>degradation</u>	Major <u>degradation</u>	Degradation of
12,000	phase	phase	sand and gravel
11,000	Aggradation of 65' terrace (dated with faunal remains)	Aggradation of 70' of valley fill (C ¹⁴ dated)	outwash to form upper terraces, Aggradation
10,000			of
9,000	<u>Degradation</u> phase	Continuous	150'
8,000	Aggradation of 20'- 30' terrace (C ¹⁴ dated and confirmed by Mazama ash marker-horizon)	phase of	of
7,000		<u>degradation</u>	valley fill
6,000		during which	(minimum age esti- mated from faunal evidence)
5,000		five unpaired	Phase of <u>degra-</u>
4,000	<u>Degradation</u> to the	terraces were	<u>dation</u> during
3,000	present channel	carved from	which several
2,000	level	the valley fill	unpaired terraces
1,000			were formed
Present			

*The time of transition between phases of aggradation and degradation is tentative.

valley of the river" (Bayrock and Hughes, 1962, p. 31). It is possible that this event may have been initiated by a decrease in discharge, since it is known that such an occurrence may change a braided stream to a meandering one (Leopold and Wolman, 1957). Further, changes in other variables, for instance sediment load, might have contributed.

At this time the Whitemud drainage system was in its formative stage of development. The basin divide already was probably determined. Modification of a preglacial valley, coincident with the southern half of the basin (Farvolden, 1963a; Carlson, 1967) took place mainly in the area west of the Edmonton International Airport. Here the deposition of till, glacio-lacustrine material, and early North Saskatchewan River alluvium obscured the preglacial thalweg. Similarly, a preglacial ridge which existed in the area of the present Whitemud Creek lower reaches was effectively covered with deposits. Thus the developing Whitemud Creek was closely adjusted to the preglacial valley centre-line in the southern half of the basin but was not in the northern sector.

The main plan features of the present drainage network were probably etched on the new depositional surfaces shortly after the North Saskatchewan channel became stabilized. Observations of developing drainage systems have shown that the greatest and most rapid changes in network configuration occur soon after the initiation of drainage. Given reasonably constant conditions of climate and crustal stability, network transformation thereafter tends to be conservative (see: Leopold, Wolman and Miller, 1964; Morisawa, 1964; Strahler, 1964).

Then came a period of maximum fluvial activity, both erosional and depositional, during which the terraces were formed. The paired

terraces of the Whitemud valley cannot be reconciled with the concept of a major valley fill being subsequently trenched to form various terrace levels. Rather, a number of degradation phases were interspersed with periods of aggradation. No meaningful absolute dates were obtained for the Whitemud deposits. A buried tree stump in the middle terrace (Section 9, discussed in Chapter Three) was sampled and subjected to C^{14} analysis.¹ The age of this wood unfortunately proved to be indistinguishable from the present. This requires explanation since the presumably related North Saskatchewan River terrace may be 11,000 years old. It appears that the tree was established on the middle terrace surface in recent times. It was then snapped off near the base, by some unknown agency, and was subsequently buried by alluvium during one or more exceptionally large recent floods. The uppermost alluvium, above the stump, is of a clay loam texture while that surrounding the wood is silty clay loam (Table XIII).

There is little reason to doubt that the Whitemud and North Saskatchewan terraces are intimately related (Figure 74). The lower Whitemud terrace grades evenly into the lowest unit recognized by Westgate (1969), and the higher ones are consistent in terms of relative elevation. No comparison of the related stratigraphies was attempted since the two rivers are of widely disparate size. If it is accepted that the Whitemud and North Saskatchewan terraces were formed contemporaneously then difficulties arise in trying to date,

¹The C^{14} analysis was carried out by Geochron Laboratories, Inc., Cambridge, Massachusetts.

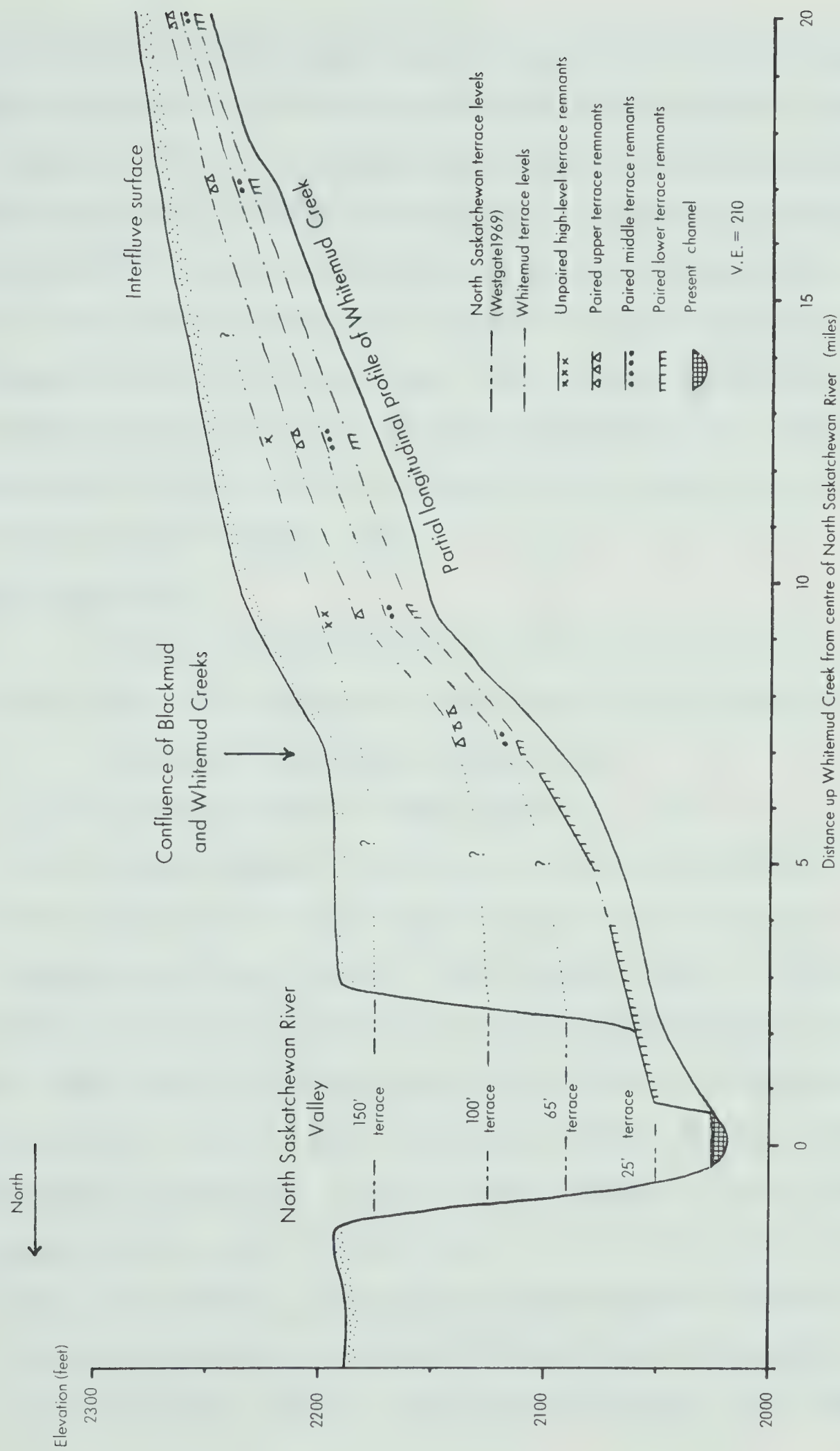


Figure 74 - North Saskatchewan River and Whitemud Creek Terrace Elevations

even generally, the Whitemud upper terrace, which is better-represented than any other. It is even more difficult to interpret the higher-level remnants, if they are representative of an earlier aggradation phase, and are not just isolated strath deposits. Lacking any evidence to the contrary, it is reasonable to conclude that the Whitemud middle terrace is in the order of 11,000 years old, that the lower terrace was formed between 8,500 and 5,000 years ago. This inference rests on the well-established dates for the lowest North Saskatchewan River terrace and Westgate's (1969) tentative correlation of the middle unit with the Cochrane member (Stalker, 1968).

4:4 Conclusion

Radiocarbon dates derived for postglacial alluvium in the Western plains (Table XVI) indicate two broad periods of net aggradation:

(1) between 11,500 and 10,000 years B.P.

and (2) between 8,500 and 4,000 years B.P.

This is in good agreement with the work of Westgate (1969) but is a little at variance with the terrace sequences at Cochrane and in the lower Red Deer valley (Stalker, 1968; McPherson, 1968). The periodic damming of rivers by changes in ice lobe and ice sheet margin positions has been cited as a probable cause of early postglacial aggradation by, among others, Stalker (1968), McPherson (1968) and Westgate (1969). Conversely, degradation has been linked to periods of rapid glacier wasting, such as the Altithermal interval. Earlier it was noted that local pollen profiles, soil types, and archaeological evidence indicate a past extension of grassland northward beyond the present limits of this vegetation complex. From evidence elsewhere it has sometimes

been assumed that this took place during the Altithermal period, approximately 8,000 to 4,000 years B.P. However, this segment of postglacial time coincides with aggradation of alluvium in the North Saskatchewan River valley. The situation is further confused when we recall the Bryson-Wendland (1968) hypothesis of an ice-free corridor opening, with a consequent drastic lowering of temperatures, about 8500 years ago. Thus, the period in which very significant changes in hydrologic regimens took place is associated with a number of conflicting hypotheses. The crux of the matter is a paucity of absolute dates and it is unlikely that many of the arguments will be resolved until much more C^{14} dating can be carried out.

It may be difficult to visualize why degradation on the grand scale envisaged by Westgate (1969), and others, took place in early postglacial time. Two arguments in support of this view can be made. First, the early postglacial period probably witnessed far greater fluctuations in climate than did the last few thousand years. Furthermore, Leopold, Wolman and Miller (1964, p. 442) noted that:

Stratigraphic studies of alluvial sequences all seem to indicate that large-scale alluviation in valley systems results from processes which act relatively slowly. But erosional features, such as valley-trenching, appear to occur more rapidly and may by the same time standards be considered episodic or catastrophic.

The second argument is that isostatic rebound following the removal of the Wisconsin ice masses was probably at a maximum soon after deglaciation.

In view of the conflicting hypotheses, hitherto outlined, and also because of the paucity of absolute dates, this interpretation of

the Whitemud basin history is rather restricted. Perhaps the most valuable observation made is that the well-developed upper terrace indicates a distinct aggradational phase long before the dated North Saskatchewan units were formed. In the latter valley, the related deposits probably were almost wholly removed during subsequent degradational periods. Further investigation of small valleys tributary to the North Saskatchewan River should reveal whether this terrace exists over a wide area.²

²This problem is being investigated by Mr. V.K. Sharma, graduate student in the Department of Geography, the University of Alberta.

CHAPTER FIVE

SUMMARY AND FUTURE RESEARCH

5:1 Summary

Within the International Hydrological Decade programme the Whitemud basin was designated as a representative unit of the Parkland zone of central Alberta. Detailed hydrological work has not yet been undertaken in this catchment, for a number of reasons, but it is probable that hydrology will receive greater attention in the near future. In view of the almost complete lack of data pertinent to the fluvial geomorphology of the basin the present investigation was designed to fulfill three needs. First, it was considered necessary to evaluate the premise that the Whitemud basin is typical of the local area. To this end morphometric methods were employed. The second objective was the measurement and description of interfluvial, valley and channel landforms characteristic of the Whitemud basin. The third aim was to present an interpretation of the geomorphic history of the basin. Together, these techniques represent a new, integrated, approach to drainage basin analysis in western Canada.

The morphometric analysis was based on the selection of all fourth order catchments within an area which minimizes variations in climate, soils, vegetation and surficial geology. Ten basins were selected. Channel numbers, dimensional properties of channel length and basin area, length, perimeter and relief, were enumerated or measured. Dimensional ratios, including measures of erosional tex-

ture, basin shape and basin relief, were also calculated for each basin. Two new methods were introduced. The first is an extension of Mueller's (1968) method of measuring channel sinuosity over selected reaches. The sinuosities of all channel segments of the given orders were derived and averaged. Comparisons of the average values led to inferences as to why channels are more sinuous in some basins than in others. The second method developed measures the channel-penetrated area of a basin. Consecutive first order channel heads were joined by straight lines to demarcate a zone of "no erosion" surrounding the drainage network. This is particularly relevant for comparing basins which have developed on continental Pleistocene deposits where networks are usually in an early stage of formation. Unusual basins can be identified and estimates of a regional central tendency may be made.

An assessment of the morphometric typicality of the Whitemud basin was achieved by ranking the deviations of variates from the associated frequency distribution modal classes. The Whitemud basin accords most closely to the local "norm" in terms of the following: the percentage of direct entrants, mean lengths of first and second order channels, mean sinuosity of third order channels, drainage density, constant of channel maintenance, all shape properties and all relief properties except maximum basin relief. Conversely, the basin is least typical with regard to the following: numbers of channels, length of the fourth order channel, average sinuosity of the fourth order channel, average sinuosity of all channels in the basin and maximum basin relief.

These morphometric parameters are of fundamental importance for any hydrological study encompassing physiographic factors. Future investigators should find the dimensionless measures of erosional texture, basin shape and basin relief of particular value for correlation with streamflow and sediment discharge variables.

The detailed description of interfluvial, valley and channel forms in the Whitemud basin serves three ends. First, the observations add to the geomorphic knowledge of the local area; second, they establish some basic quantitative measures of hydrological significance (for example, channel slope and channel cross-section form); and third, they form a base for an interpretation of the geomorphic history of the basin.

Interfluvial maximum slope angles are generally flat. Exceptions occur mainly near the basin divide and where the surficial geology is non-lacustrine. However, there appears to be little difference between the mean maximum slope angles of the lacustrine and non-lacustrine zones. The preferred orientations of micro-slopes on the interfluvial surfaces fall in the north-east and south-west sectors. This may indicate partial control of the interfluvial slopes by underlying aligned till ridges of one sort or another.

The main Whitemud valley contains abundant evidence of three paired terrace levels. These stand about 10, 20 and 40 feet, respectively, above the present channel near the basin mouth. With increasing distance up-valley these merge with the present channel. Remnants of a possible higher aggradation surface exist.

The average textures of the middle and upper terraces are both

categorized as loam on the basis of sample grain-size analysis. The lower terrace is comparatively rich in sand and has an average texture transitory between loam and sandy loam. In terms of gross structure the sedimentary sequences are comparable. A basal unit of gravel usually overlies bedrock and is in turn overlain by fine-grained alluvium. The fine-grained alluvium of the middle and upper terraces differs slightly from that of the lower terrace in that a silt and clay-rich stratum often succeeds the basal gravel. Above this layer slightly sandier material occurs. This textural arrangement differs from the usual fluvial sedimentary sequence inasmuch as the silty and sandy beds are reversed. Possibly the upper sandy material was deposited by abnormally large floods near the end of the related aggradational phase, or early in the succeeding degradational period. The relative homogeneity of the lower terrace fine-grained alluvium suggests a more constant hydraulic regime during aggradation and throughout the last phase of degradation. But, occasional extreme floods have occurred in recent years because at least one middle terrace site exhibits evidence of very recent aggradation, 15 feet above the present channel.

The valley-wall slope angles reflect the rejuvenative history of the Whitemud basin. These slopes are steepest near the basin mouth and decrease systematically up-valley for several miles but there is no evidence of significant valley asymmetry; average slopes of the older parts of the walls may have declined approximately 3 degrees since their initial formation.

The average cross-section form of the main Whitemud channel is

shallow and wide compared to cross-sections of the remaining portion of the fourth order channel and of the third order channels. While no one variable can wholly explain this difference, variations in bed resistance may be critical. A continuous veneer of gravel in the main Whitemud channel segment acts as a partial armour against bed erosion. The remaining channel segments, lacking this veneer, are incised in relatively homogeneous glacial and glacio-lacustrine deposits. Bed resistance is, therefore, less pronounced, and channels are deeper and narrower. The partial longitudinal profile of the Whitemud Creek illustrates the effects of both rejuvenation and structural control. Two marked breaks in slope reflect lithologic boundaries but the general convex-upward form of the profile is a function mainly of continued rejuvenation.

The grain-size distribution of riffle debris is determined largely by the local geological environment. While riffle-formation is directly attributable to hydraulic factors, the source of bed load sediments is critical in determining the grain-size characteristics. Where gravel is readily available (from relatively indurated bedrock strata, alluvial terrace gravels and till beds) the channel bed, in both pools and riffles, has a gravel lag-deposit. Reworking of this material by the stream creates gravel riffles. The average size of the riffle gravel varies little with distance upstream. Very large sandstone slabs at one site remain close to the bedrock source and have not been re-located more than 50 yards downstream. It is concluded that the relatively homogeneous size of riffle gravel along the main segment of the fourth order channel is a result of sorting over a number of years

under a comparatively constant hydraulic regime. In those channel reaches where gravel is not readily available the riffles are formed of fine-grained sediments. A few stones remain trapped in pools but the channel beds are dominated by sand, silt and clay. The transition from gravel to fine-grained riffles is abrupt and coincides with lithologic changes where the fourth order channel is first incised completely through the glacial and glacio-lacustrine deposits into bedrock.

An interpretation of the Whitemud basin geomorphic history is dependent on the paired terrace sequence in the main valley. Although one attempt to obtain a meaningful C^{14} date for material from the middle Whitemud terrace was unsuccessful, a North Saskatchewan River terrace which grades evenly into the lower Whitemud Creek terrace is dated at 8,500 to 5,000 years B.P. A tentative correlation of the three (or possibly four) Whitemud Creek terraces with the four North Saskatchewan River terraces is suggested. A more definitive assessment of the local geomorphic history awaits detailed geochronological work on the higher terraces.

5:2 Future Research

In much geomorphic and hydrological work basins for study are often chosen rather arbitrarily. A more objective means of basin selection is the determination of central tendency values of morphometric parameters for a region. The "most typical" catchment can then be chosen. Pilot-studies carried out for this purpose need not include all the morphometric techniques outlined here. The measures adopted will depend on the initial aims of each project.

Local geomorphic problems requiring intensive research are numerous but four need special attention. First, the analysis of valley-wall slopes should be extended to other valleys where micro-climatic variations are significant. The conclusion that slope decline has occurred in parts of the Whitemud system is tentative and requires further testing. This evidence is important in that it supports the inference, from other sources, that the uppermost alluvial terraces were formed in early Holocene time.

The second, and perhaps most important problem, concerns post-glacial alluvial terraces in central Alberta. Only through detailed mapping, the analysis of textural and structural characteristics, and especially through C^{14} dating of palaeols, bones and other organic material, can a satisfactory postglacial chronology be constructed. This is especially true of the older, higher terrace deposits.

The third possible topic outlined here belongs in part to historical geography. An analysis of land-use changes since the time of the first European settlement could point to geomorphological and hydrological effects of such factors as the disruption of natural vegetation and the building of roadside ditches. This entails the study of exploration reports, old photographs and newspapers, as well as interviews with surviving pioneers.

Finally, it is reiterated that current geomorphic processes active here are poorly-known. Although many difficulties exist, not least being the generally slow rates of erosion in this environment, almost any study of process is better than none. Of necessity,

short-term field studies of erosion and deposition should be concentrated in areas where processes are relatively active. While this type of approach yields little information about average regional rates of action it may at least provide knowledge of one end of the process-spectrum.

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APPENDIX A

DEVIATIONS OF MORPHOMETRIC VARIATES FROM
THE FREQUENCY DISTRIBUTION MODAL CLASSES

Basin - Whitemud		Mode								
Variate		-4	-3	-2	-1	0	+1	+2	+3	+4
N_1										x
N_2									x	
N_3									x	
N_k									x	
% E						x				
$\bar{L}_1$						x				
$\bar{L}_2$						x				
$\bar{L}_3$							x			
$\bar{L}_4$										x
$\bar{L}_k$							x			
$\bar{G}_1$								x		
$\bar{G}_2$					x					
$\bar{G}_3$						x				
$\bar{G}_4$										x
$\bar{G}_k$									x	
A										x
D						x				
C						x				
F							x			
F_1							x			
D_1				x						
S_1						x				
S_2						x*				
S_3						x				
S_4						x				
k						x*				
K						x				
H									x	
H_d						x*				
R_1						x				
R_2						x				
R_3						x				
Comparative Score		0	0	1	1	16	4	1	5	4

* Bi-modal frequency distribution

Basin - Pipestone

Variate	-4	-3	-2	-1	0	+1	+2	+3	+4
N_1					x				
N_2				x					
N_3				x					
N_k				x					
% E			x						
$\bar{L}_1$								x	
$\bar{L}_2$							x		
$\bar{L}_3$									x
$\bar{L}_4$					x				
$\bar{L}_k$									x
$\bar{G}_1$							x		
$\bar{G}_2$					x				
$\bar{G}_3$					x				
$\bar{G}_4$					x				
$\bar{G}_k$					x				
A								x	
D			x						
C								x	
F				x					
F_1				x					
D_1	x								
S_1					x				
S_2					x*				
S_3					x				
S_4					x				
k				x*					
K					x				
H									x
H_d					x*				
R_1					x				
R_2						x			
R_3						x			
Comparative Score	1	0	2	6	13	2	2	3	3

*Bi-modal frequency distribution

Variate	-4	-3	-2	-1	0	+1	+2	+3	+4
N_1					x				
N_2				x					
N_3				x					
N_k				x					
% E			x						
$\bar{L}_1$					x				
$\bar{L}_2$						x			
$\bar{L}_3$					x				
$\bar{L}_4$					x				
$\bar{L}_k$						x			
$\bar{G}_1$						x			
$\bar{G}_2$						x			
$\bar{G}_3$					x				
$\bar{G}_4$					x				
$\bar{G}_k$					x				
A					x				
D			x						
C							x		
F					x				
F_1					x				
D_1	x								
S_1						x			
S_2							x*		
S_3							x		
S_4					x				
k				x*					
K					x				
H							x		
H_d					x*				
R_1						x			
R_2					x				
R_3									x
Comparative Score	1	0	2	4	14	6	4	0	1

*Bi-modal frequency distribution

Basin - South Pipestone Tributary

Variate	-4	-3	-2	-1	0	+1	+2	+3	+4
N_1					x				
N_2					x				
N_3					x				
N_k					x				
% E				x					
$\bar{L}_1$				x					
$\bar{L}_2$								x	
$\bar{L}_3$					x				
$\bar{L}_4$						x			
$\bar{L}_k$						x			
$\bar{G}_1$					x				
$\bar{G}_2$				x					
$\bar{G}_3$					x				
$\bar{G}_4$							x		
$\bar{G}_k$						x			
A						x			
D				x					
C						x			
F					x				
F_1					x				
D_1	x								
S_1							x		
S_2						x*			
S_3									x
S_4							x		
k					x*				
K			x						
H								x	
H_d					x				
R_1							x		
R_2								x	
R_3									x
Comparative Score	1	0	1	4	11	6	4	3	2

*Bi-modal frequency distribution

Basin - Blackmud

Variate	-4	-3	-2	-1	0	+1	+2	+3	+4
N_1						x			
N_2					x				
N_3					x				
N_k					x				
% E					x				
$\bar{L}_1$					x				
$\bar{L}_2$					x				
$\bar{L}_3$					x				
$\bar{L}_4$					x				
$\bar{L}_k$					x				
$\bar{G}_1$					x				
$\bar{G}_2$						x			
$\bar{G}_3$					x				
$\bar{G}_4$					x				
$\bar{G}_k$				x					
A					x				
D					x				
C					x				
F							x		
F_1							x		
D_1					x				
S_1								x	
S_2					x*				
S_3								x	
S_4								x	
k					x*				
K			x						
H					x				
H_d					x*				
R_1					x				
R_2						x			
R_3					x				
Comparative Score	0	0	1	1	22	3	2	3	0

*Bi-modal frequency distribution

Variate	-4	-3	-2	-1	0	+1	+2	+3	+4
N_1							x		
N_2					x				
N_3						x			
N_k						x			
% E						x			
$\bar{L}_1$					x				
$\bar{L}_2$						x			
$\bar{L}_3$						x			
$\bar{L}_4$					x				
$\bar{L}_k$					x				
$\bar{G}_1$					x				
$\bar{G}_2$					x				
$\bar{G}_3$					x				
$\bar{G}_4$						x			
$\bar{G}_k$					x				
A							x		
D				x					
C						x			
F					x				
F_1					x				
D_1		x							
S_1						x			
S_2							x*		
S_3									x
S_4							x		
k						x*			
K				x					
H					x				
H_d						x*			
R_1					x				
R_2					x				
R_3					x				
Comparative Score	0	1	0	2	14	10	4	0	1

*Bi-modal frequency distribution

Variate	-4	-3	-2	-1	0	+1	+2	+3	+4
N_1					x				
N_2				x					
N_3					x				
N_k				x					
% E		x							
$\bar{L}_1$					x				
$\bar{L}_2$					x				
$\bar{L}_3$					x				
$\bar{L}_4$					x				
$\bar{L}_k$					x				
$\bar{G}_1$							x		
$\bar{G}_2$					x				
$\bar{G}_3$					x				
$\bar{G}_4$						x			
$\bar{G}_k$						x			
A					x				
D				x					
C						x			
F					x				
F_1					x				
D_1		x							
S_1					x				
S_2					x*				
S_3					x				
S_4					x				
k					x*				
K					x				
H					x				
H_d					x*				
R_1					x				
R_2					x				
R_3					x				
Comparative Score	0	2	0	3	23	3	1	0	0

*Bi-modal frequency distribution

Variate	-4	-3	-2	-1	0	+1	+2	+3	+4
N_1						x			
N_2					x				
N_3					x				
N_k					x				
% E				x					
$\bar{L}_1$					x				
$\bar{L}_2$					x				
$\bar{L}_3$					x				
$\bar{L}_4$						x			
$\bar{L}_k$					x				
$\bar{G}_1$						x			
$\bar{G}_2$						x			
$\bar{G}_3$					x				
$\bar{G}_4$							x		
$\bar{G}_k$						x			
A					x				
D					x				
C					x				
F						x			
F_1						x			
D_1					x				
S_1									x
S_2					x*				
S_3							x		
S_4							x		
k						x*			
K				x					
H					x				
H_d				x*					
R_1					x				
R_2				x					
R_3					x				
Comparative Score	0	0	0	4	16	8	3	0	1

*Bi-modal frequency distribution

Variate	-4	-3	-2	-1	0	+1	+2	+3	+4
N_1							x		
N_2						x			
N_3							x		
N_k						x			
%E					x				
$\bar{L}_1$				x					
$\bar{L}_2$					x				
$\bar{L}_3$					x				
$\bar{L}_4$					x				
$\bar{L}_k$					x				
$\bar{G}_1$					x				
$\bar{G}_2$					x				
$\bar{G}_3$							x		
$\bar{G}_4$									x
$\bar{G}_k$							x		
A					x				
D							x		
C				x					
F								x	
F_1								x	
D_1					x				
S_1								x	
S_2					x*				
S_3						x			
S_4									x
k					x*				
K	x								
H						x			
H_d						x*			
R_1								x	
R_2								x	
R_3							x		
Comparative Score	1	0	0	2	11	5	6	5	2

*Bi-modal frequency distribution

Basin - Bigstone

Variate	-4	-3	-2	-1	0	+1	+2	+3	+4
N_1						x			
N_2					x				
N_3					x				
N_k					x				
% E					x				
$\bar{L}_1$						x			
$\bar{L}_2$					x				
$\bar{L}_3$						x			
$\bar{L}_4$					x				
$\bar{L}_k$						x			
$\bar{G}_1$							x		
$\bar{G}_2$				x					
$\bar{G}_3$							x		
$\bar{G}_4$					x				
$\bar{G}_k$					x				
A					x				
D					x				
C					x				
F						x			
F_1						x			
D_1					x				
S_1					x				
S_2				x*					
S_3						x			
S_4					x				
k				x*					
K					x				
H								x	
H_d				x*					
R_1						x			
R_2						x			
R_3								x	
Comparative Score	0	0	0	4	15	9	2	2	0

*Bi-modal frequency distribution

APPENDIX B

LABORATORY TECHNIQUE FOR MECHANICAL GRAIN-SIZE ANALYSIS*

- (1) Air-dry samples in open trays.
- (2) Pass sufficient material for test through 2mm. sieve.
- (3) From 2mm. sieved material weigh out two 10gm. samples (weight accurate to nearest mg.).
- (4) Wash both samples into 600ml. beakers with distilled water. Add sufficient water to bring volume of solution up to about 50ml.
- (5) Add to each beaker 5ml. of H_2O_2 and place on hot plate (set to 150°F) until frothing ceases. When frothing ceases add H_2O_2 dropwise until no further frothing occurs. Boil solution to remove excess H_2O_2 .
- (6) Clean sides of beaker with gendarme, washing residue into solution.
- (7) Contents of one beaker are transferred to evaporating dish and dried overnight at 105°C. This is subsequently cooled in a dessicator and weighed to obtain BASIC WEIGHT (weight of sample minus water and organic matter).
- (8) Add 10ml. of "Calgon" plus Na_2CO_3 solution to the second beaker and allow to stand overnight.
- (9) Solution from second beaker is transferred to stirring-cup and sufficient water to half-fill the cup is added. The solution is then dispersed for 10 minutes with the electric-disperser.
- (10) Transfer this solution to a graduated cylinder and top up with water to 1000ml.
- (11) Stir thoroughly with rod for 1 minute. Place cylinder in constant temperature (25°C) water-bath and record time.
- (12) After 7 hours remove 10ml. pipette aliquot from each cylinder, at a depth of 10cm. Transfer each aliquot to an evaporating dish and dry overnight (SAMPLE A).
- (13) Pour off supernatant fluid from each cylinder. Wash contents into 600ml. beaker marked at 10cm. above base.
- (14) Add cold tap-water to bring solution up to 10cm. mark. After 44 seconds decant the supernatant fluid through a 270 mesh

sieve. Repeat this procedure until supernatant fluid is clear after 44 seconds. Transfer sand to sieve and wash thoroughly. Transfer sand from sieve to an evaporating dish and dry overnight (SAMPLE B).

- (15) Weigh dried samples after cooling in dessicator. From SAMPLE A calculate clay as a percentage of the BASIC WEIGHT (subtracting 0.004mg. for weight of dispersing agent). Calculate sand as a percentage of the BASIC WEIGHT. Obtain the silt percentage by subtraction.

*Laboratory procedure outlined by Dr. R.B. Bryan, Department of Geography, University of Alberta.

APPENDIX C

COMPUTING FORMULAS FOR THE CHI-SQUARE, HOMOGENEITY OF VARIANCES AND
DIFFERENCE OF MEANS TESTS*(1) Chi-square (χ^2)

$$\chi^2 = \sum \frac{(fo - fe)^2}{fe}$$

where fo = observed frequencies
and fe = expected frequencies

and $\chi_c^2 = \chi_{\gamma, \alpha}^2$

where γ = degrees of freedom
= $n - 1$
and α = level of significance

(2) Homogeneity of variance (F test)

$$F = \frac{s_1^2}{s_2^2}$$

where s_1^2 and s_2^2 are the sample
variances (s_1^2 is always taken as
the larger variance).

and $F_c = F_{\alpha, \gamma_1, (n-1), \gamma_2, (n-1)}$

where α = level of significance
 γ_1 = degrees of freedom of the
numerator
 γ_2 = degrees of freedom of
denominator
and n = sample size

(3) Difference of means (Student t test)

$$t = \frac{\bar{x}_1 - \bar{x}_2}{\sqrt{\frac{(n_1-1)s_1^2 + (n_2-1)s_2^2}{n_1 + n_2 - 2} \left(\frac{1}{n_1} + \frac{1}{n_2}\right)}}$$

and $t_c = t_{\gamma, \alpha}$

where $\bar{x}_1$ and $\bar{x}_2$ are the
sample means (other symbols
as previously defined)

and $\gamma = n_1 + n_2 - 2$

*See: Blalock, 1960; Cole and King, 1968; Croxton and Cowden, 1955.

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